MAR 2 1 1916

VOLUME XXIII NUMBER 1

THE

JOURNAL OF GEOLOGY

JANUARY-FEBRUARY 1915

THE ORIGIN OF THE INCLUSIONS IN DIKES

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PART I

INTRODUCTION

Several dikes containing numerous inclusions have recently been seen by the writer, and the origin of these inclusions appears to be of sufficient interest to warrant a brief notice of these and other cases found in the literature.[1] The dikes have received various descriptive names, but none of them is sufficiently comprehensive to include all the examples.

Inclusions are not infrequent in all types of both extrusive and intrusive rocks. Those in extrusive rocks have been specially treated by Lacroix. Intrusive rocks are so extensive and of such varying forms that only dikes and a few related intrusions will be considered here. The xenoliths in the large intrusive bodies have formed the basis for the stoping hypothesis, but only part of the inclusions in dikes are of this origin.

Dikes acquire inclusions by shattering blocks off the walls of a fissure during their ascent through it. These fragments may remain near the place from which they came, or they may move up or down. In most cases the fragments rise, whether of greater specific gravity than the molten dike-rock or not, because they are forced upward by the magma. In the cases where the fragments sink, either they are heavier than the magma or they are carried down.

Some dikes have invaded conglomerates, from which they have dissolved the cement and included the bowlders. In other cases the fissures through which the magma came may have been open to the surface, so that stream gravels fell in and were caught in the ascending magma. In rare instances a dike may ascend through a fault breccia and thus acquire its inclusions. These special cases are not separated from the other examples.

To classify dikes according to the direction of movement of the inclusions is difficult, because in almost every case some inclusions rise and some remain near the place of origin. A classification is attempted which will at least serve to emphasize the number of examples in which inclusions sink.

[1] The writer is indebted to Professor R. A. Daly for suggestions concerning this paper.

EXAMPLES

(*A*) *All inclusions near place of origin.*—

Cornwall, England: In the southern part of Cornwall, near Carrick Luz, there is a gabbro boss surrounded by a fringe of gabbro dikes extending outward through the serpentines of the region. All of the dikes contain numerous inclusions of the serpentine. These dikes also illustrate injection foliation.[1]

Da la Bèche reported one of the granite dikes of this region to contain numerous fragments of the slate which it cuts.[2]

Mexico: In the state of Guerrero, Mexico, a granite sill from 12 to 15 feet wide cuts the Juratrias shales, and contains a number of fragments of the slate a foot in diameter. The vertical sill runs parallel to the bedding of the shales, but has evidently broken off the shale inclusions from the walls near the place where they are frozen in the dike. The inclusions are now nearly weathered out of the granite, so that the effect of the heating is indeterminate, but the granite adjoining them shows no immediate contact effect.[3]

Cape Ann, Massachusetts: At Pigeon Cove, Cape Ann, is a labradorite porphyry dike of very coarse grain, 25 feet wide and traceable for 3 miles, everywhere cutting alkaline hornblende granite, but including fragments of diabase and quartzite. The inclusions are about 6 inches in diameter and have a subangular to rounded outline. The numerous diabase dikes now exposed in the region cut the porphyry dikes. The inclusions must have come from quartzites and diabase dikes cutting them, forming the roof of the granite batholith, and must have sunk or been carried down in the dike to their present level. Since that time, which was probably the Carboniferous, the quartzites from the roof of the batholith and the upper part of the batholith itself have been completely removed by erosion. Quartzites appear in small amounts elsewhere in Essex County. The distance through which the inclusions must have

[1] J. S. Flett, "The Geology of the Lizard and Meneage," *Mem. Geol. Surv. Great Britain*, 1912; also, *Proc. Geologists' Association*, XXIV (1913), 127.

[2] *Geological Report on Cornwall* (1839), p. 182.

[3] The writer is indebted to Mr. Y. S. Bonillas for this information.

descended can only be estimated as probably several thousand feet.[1]

(*B*) *Some inclusions have sunk.*—

Montreal, Canada: A number of breccia dikes have been described from the vicinity of Montreal and the Monteregian Hills by Robert Harvie.[2] The dikes will be mentioned in the order in which they are described by Harvie.

Near La Trappe in the Oka Mountains are two inclusion-bearing alnöite dikes, one of which is enlarged to a width of 180 feet, while the other is 25 feet wide and traceable for half a mile. The fragments are of Grenville limestone, Laurentian gneiss, Potsdam sandstone, and Beekmantown sandstone. The last two formations are stratigraphically higher than the exposure which is near the contact of the Grenville and Laurentian. Therefore some of the blocks rose, others sank.

Near Ste. Anne de Bellevue is a dike a foot wide cutting Trenton limestone and containing angular fragments of sandstone, hornstone, and limestone. The sandstone has a quartzitic rim, the hornstone was apparently originally a shale, and the limestone is changed to a crystalline marble.

On Westmount Mountain at Little's quarry is a large breccia camptonite dike about 4 feet wide. In one side of the dike an offshoot from it "has eroded or stoped out and filled a large cavity." The heat of the intrusive has baked the limestone, and a section from the unaltered limestone to the dike shows increasing baking and shattering until the blocks of limestone drop off into the dike-rock, where they form the bulk of the angular inclusions. Harvie says:

> In this process of shattering by which the action of stoping went forward every gradation is visible, from the solid unaltered limestone to the fragments finally frayed off and held in suspension in the breccia. The circulation of the materials of the intrusive has evidently been very vigorous, since fragments of other rocks, granites, etc., are found carried up to the top of the stope within a foot of the unshattered limestone. The magma has not exerted any pro-

[1] The writer is indebted to Professor C. H. Warren, of Boston, for the information concerning the inclusions in this dike.

[2] On the origin and relations of the Paleozoic breccia in the vicinity of Montreal, see *Trans. Roy. Soc. Canada*, 3d ser., III (1909), 249–78.

nounced solvent action on the inclusions, either the basic limestone or the more acid granites; neither has the heat greatly affected the limestone, which one would expect to find in a crystalline state.

There are a number of other breccia dikes in this immediate vicinity, 2 to 7 feet wide, of both a camptonite and a pyroxenite composition. They contain fragments of granite, gneiss, essexite, syenite, Potsdam sandstone, and Trenton limestone, the country rock. All these inclusions have either come from below or from the limestone near by.

It is evident that the inclusions in the cases above mentioned have been derived almost wholly by the dike-magma shattering the walls as it advanced. There must have been an active circulation in the dikes to keep them at a high temperature at the top while they made their way through the limestone and probably through the underlying rocks. This circulation carried up the blocks of the pre-Cambrian and carried down the blocks of the Potsdam and Beekmantown. These dikes appear, therefore, to be cases where the ascent was due, at least partly, to a blowpiping action on the rock overhead.

In the same memoir Harvie describes other dikes and intrusions of various forms, all of which consist of a breccia of similar inclusions in a basic matrix. One of the masses of breccia on St. Helen's Island,[1] near Montreal, cutting Utica shale, contains blocks of fossiliferous Oriskany limestone and Helderberg limestone, indicating the former presence of strata of these ages in the region. It also contains fragments of the Ordovician and older rocks.

The vertical movement of the fragments in these cases is interesting because it has been in both directions in the same dike or mass. This movement does not seem to be dependent on the form of the intrusion nor on the composition of the intruding dike-rock. The thickness of the formations from the top of the Laurentian to the bottom of the Utica is 2,500 feet (Harvie). The Utica is succeeded by Lorraine shales 2,000 feet thick. No higher stratigraphic horizon is exposed in the region, except in the inclusions on St. Helen's Island. Therefore these Devonian fragments must

[1] See also F. D. Adams, *Twelfth Internat. Geol. Cong., Guide Book* 3, p. 55, for a description of these breccias.

have sunk or been carried down over 2,000 feet, and the Laurentian fragments in the same breccia brought up over 2,500 feet. In the La Trappe dike the movement in each direction was probably not as great, but the distance from the Beekmantown down to the contact of the Grenville and Laurentian must have been several thousand feet.

Marblehead, Massachusetts: At Marblehead, Massachusetts, half-way between Peaches Point and Naugus Head, are a number of dikes of pulaskite cutting an igneous complex. These dikes range in width from a few inches to 50 feet. In many cases they appear to follow the course of earlier diabase dikes, all stages being present, from a network of pulaskite veins in a solid diabase dike to a shatter breccia of diabase in pulaskite and finally to a band of schistose streaks of diabase in the center of a broad dike of pulaskite, showing the former position of a diabase dike. In one case there are remnants of two such dikes in a wide pulaskite dike. The diabase fragments are lenticular schistose bands, sometimes several inches in width, in a coarse-grained pulaskite, which grows finer-grained in the small tongues which permeate the diabase inclusions. The dark constituents of the pulaskite appear to have been derived largely from the absorbed diabase. The width of the original dikes was from 1 to 3 feet.

Southern Sweden: At Brevik and at Karlshamn, southern Sweden, diabase dikes filled with rounded inclusions occur. They have been described by Hedström[1] and Eichstädt[2] on whose papers the following description is based. Two of these dikes are known, being mapped on the Eskjö and Karlshamn sheets of the Swedish Geological Survey.

The dike at Karlshamn is 3 miles long and in places several hundred feet wide. The inclusions are all on the west side of the dike and consist wholly of quartzite and quartzitic sandstone. Cataclastic structure is characteristic of all the inclusions, but this texture was caused by metamorphism before the quartzites were

[1] "The Pebble-Diabase of Brevik," *Eleventh Internat. Geol. Cong., Guide Book 18*, 1907, pp. 47–51.

[2] "Om quartsit-diabas-konglomeratet i Småland och Skåne," *Sveriges Geol. Unters.*, Ser. c, No. 74, 1885.

included in the dike. The contact of the pebbles and the diabase in the case of this dike is sometimes sharp and sometimes indistinct. In the pebbles, which have undergone the most marked absorption, it is difficult to distinguish even the center of the pebble. Sometimes ilmenite and fine feldspar crystals have formed in the pebbles with a rim of epidote and chlorite between the pebble and the diabase. The dike cuts Archean gneiss, and none of the younger pre-Cambrian sedimentary series is present in the region.

The "pebble-diabase" dike of Brevik (on the Eksjö sheet) has a strike of about N.–S. and has been traced for about 15 miles. It cuts Almesåkra pre-Cambrian sandstone, quartzite, and conglomerate. The width of the dike varies in different exposures and it often exceeds 300 feet. Inclusions appear only in some of the outcrops, and there they are, as a rule, confined to smaller zones, often 10 to 15 feet in width, which are elongated parallel to the sides of the dike. These zones may be situated either on the sides of the dike, and usually on the eastern side, or near the middle. They vary in width from 4 to 50 feet, changing with the width of the dike. In smaller offshoots from the dike, about 15 feet in width, fragments sometimes occur evenly distributed over the entire width. Elsewhere there is a sharp boundary between the parts free from and full of inclusions.

The inclusions consist largely of quartzite and schist derived from the Almesåkra pre-Cambrian complex, with some granites, leptites, gneisses, and other pre-Cambrian rocks. The size of the fragments varies from a few inches to 30 feet. The shape of the quartzite inclusions is rounded, of the granite, subangular, while the inclosures of schist form thin bands of considerable length, which are surrounded by sheets of diabase. The pebbles are so numerous as almost to touch. They occupy about half the volume of the dike, and at times even more. The inclusions weather out easily.

In the pebble-bearing parts of the diabase the ophitic diabasic structure is not developed, and the presence of the numerous inclusions has caused a segregation of the light and dark minerals into separate spots. There are also grains of quartz and feldspar scattered through the diabase as if derived from the resorption of the edges of some of the inclusions or from the cement of the original

conglomerate from which the pebbles were derived (Hedström's theory). The inclusions of quartzite are generally surrounded by zones in which iron ore, biotite, and chlorite are developed. The contact of the diabase with the quartzite pebbles is, however, usually sharp, while the contact with the sandstone and granite is more or less indistinct, the magma having formed fused contacts and even having penetrated a few of the inclusions. In places a flow structure is developed in the diabase around the fragments.

Eichstädt proposes the theory that the inclusions in both dikes were derived from loose pebbles on the surface of the ground, probably accumulated in valleys, which fell into fissures formed in advance of the diabasic intrusions and were caught in the magma when it ascended in the fissures. In support of this theory he points out that if the blocks of quartzite were originally angular and the edges resorbed where they are now, the silica content of the diabase near the zones of inclusions would be much greater than elsewhere in the dike. This is not the case, although he finds free quartz and micropegmatite present in the diabase of both dikes. He noted the existence of pre-Cambrian conglomerate in the vicinity, but considers that the pebbles were derived from a loosely consolidated conglomerate (as this Almesåkra conglomerate may have been in the pre-Cambrian time when the dikes were formed) and not from a massive conglomerate.

Hedström proposes a somewhat similar theory: that larger and smaller pieces of the Almesåkra conglomerate have been imbedded in the diabase magma. He considers that the conglomerate was at least somewhat consolidated and accounts for part of the matrix by the numerous grains of quartz and feldspar scattered in the diabase. In support of this view he finds inclusions to which portions of the original matrix are still attached. The hard quartzite pebbles in the Almesåkra conglomerate are greatly cracked and show the same cataclastic structure as do those of the inclusions which have often split up into several pieces which have been more or less widely displaced from each other in the magma. The cement of the original conglomerate was gritty, loose, and brittle, and therefore its cohesion was easily destroyed at the intrusion of the diabase. "In the appearance and character of the pebbles, as

also in the proportions of the rock types represented among them, is a perfect agreement between the Almesåkra conglomerate and the pebble-diabase." It should be noted, however, that the Almesåkra series is now present only near Brevik. A former extension of the series for 90 miles south, to Karlshamn, may have existed.

Cripple Creek, Colorado: At Cripple Creek a number of phonolite and rhyolite dikes include or are capped by loose conglomerate and volcanic breccia. These phenomena have been described by G. H. Stone[1] who has confounded the origin of the conglomerate and breccias with that of the dikes. As pointed out by Ransome and Lindgren,[2] the former are stream gravels and volcanic breccias, some of which have been invaded by dikes. It appears probable that the inclusions in the dikes themselves have fallen in from the overlying loosely consolidated beds.

At Grizzly Peak, Colorado, similar granitic breccias have been reported by G. H. Stone[3] and it is probable that here also stream gravels or volcanic breccias have been invaded by dikes.

The manner in which an igneous rock invades a conglomerate is well illustrated in a satellitic stock of the Bayonne batholith in the Selkirk Mountains, British Columbia, as described by R. A. Daly.[4] The granitic magma has eaten its way into the conglomerate,

> dissolving out the cement in large amount, and has thus not only thoroughly impregnated the conglomerate with the granitic material, but has quite separated many of the larger quartzitic pebbles, which, still rounded, are now completely inclosed in granite. The cement was evidently more soluble in the magma than were the quartzite pebbles—a conclusion to be expected in view of the fact that the heterogeneous cement has a lower fusion-point, and in relation to the acid granite, a lower solution-point of temperature, than the more highly siliceous quartzite. The partial absorption of the conglomerate must have taken place when the magma was (because cooled down) sufficiently viscous to allow of the suspension of the blocks and pebbles. At an earlier period, when the cooling was less advanced, the quartzite pebbles

[1] "The Granitic breccias of the Cripple Creek Region," *Amer. Jour. Sci.*, Ser. 4, V (1898), 21–32.

[2] *U.S. Geol. Surv., Professional Paper 54.*

[3] *Amer. Jour. Sci.*, Ser. 4, VII (1899), 184–86.

[4] "The Geology of the N. A. Cordillera at the 49th Parallel," *Can. Geol. Surv., Memoir 38* (1914), p. 300.

themselves, like the main quartzitic and schistose formations, could have been dissolved.

Pequawket Mountain, New Hampshire: Pequawket Mountain, the Eastern Kearsarge of New Hampshire, and Moat Mountain, near by, are composed of masses of quartz porphyry, of stock-like form, which contain so many fragments of metamorphic rocks that they have been described as breccias.[1] The Pequawket mass is about 1,200 feet long and 450 feet wide. It lies at the contact of the older Albany granite and slate. The angular inclusions are very numerous, consisting largely of slates, sandstones, and phyllites, derived from the adjacent terranes. The inclusions do not show any alteration at their contact with the porphyry. The porphyry matrix is vitrophyric in the Pequawket mass, granophyric in the Moat mass. As the inclusions do not show as much metamorphism as the surrounding slates, and as the quartz porphyry is not squeezed, Daly concludes that the inclusions have probably come from above the present exposure, where the metamorphism was not so great.

Lages, Brazil: Near Lages, Brazil, an inclusion-bearing dike was found by Woodworth.[2]

The dike is of trap and was evidently the feeder of one or more of the Triassic trap sheets which are exposed in the Lages area. At the outcrop investigated, the width of the dike is 75 feet. Here the inclusions of foreign rocks constitute one-half of the volume of the dike and comprise red and black shale, coarse-grained basalt, fine-grained basic rock, and amygdaloidal basalt. The fragments of sedimentary rock were apparently disrupted from the walls of the fissure which appears to have been a fault-line. Whether they came from above or below the exposure could not be determined. The amygdaloidal basalt fragments must have come from the overlying Triassic flows according to the determination of the geological structure by Woodworth and others. Furthermore, it is certain that the dike was not formed until after other lavas had been extravasated and cooled to yield the inclusions which sank in the fluid dike-magma.

[1] R. A. Daly, *Science*, N.S., III (1896), 752.

[2] J. B. Woodworth, "Geological Expedition to Brazil and Chile. 1908–9," *Bull. Mus. Camp. Zoöl., Harvard*, LXI, No. 1, p. 95.

THE MIDDLE AND UPPER DEVONIAN OF THE ROMNEY, WEST VIRGINIA, REGION[1]

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INTRODUCTION

The name Romney shales, from "Romney in West Virginia," was published by N. H. Darton in 1892 "for the basal series of dark shales [in the Devonian]."[2] The formation was briefly described as follows:

> The basal members [of the Devonian] are fissile shales, in greater part black or dark brown in color, containing occasional thin beds of sandstone and limestone. Their average thickness is about 600 feet. . . . . The Devonian formations are not fossiliferous at many horizons in the region west of Staunton. In the Romney shales the following species are Corniferous [misprint for conspicuous]: *Discina lodensis, D. minuta, Orthis leucosia, Strophodonta demissa, Cyrtina hamiltonensis, Spirifera mucronatus, S. granulifera,* and *Leiorhynchus limitaris.* This is a Hamilton group fauna, but the stratigraphic range of Hamilton group equivalents in the Romney shales is not apparent, and Hamilton deposits probably extend some distance above.[3]

[1] Published by permission of Dr. William Bullock Clark, state geologist of Maryland.

[2] *American Geologist,* X (July, 1892), 17, and the name first appears in the table of formations on p. 13.

[3] *Ibid.,* pp. 17, 18.

The Staunton folio by Darton, which appeared in 1894, was the first one published by the United States Geological Survey for the Appalachian part of the Virginias. In this folio Darton gave the following description of the Romney shales:

> The rocks consist of dark shales, black and fissile below, but somewhat lighter and more compact above. Some of the basal beds are carbonaceous to a moderate degree, and they have been worked at several points with the mistaken idea that they might prove to be coal-bearing. The formation includes occasional calcareous streaks not far from its base, and the upper members contain alternations of thin, pale-brown or dark-buff sandy beds, which constitute beds of passage into the next succeeding formation. The vertical range and stratigraphic position of these passage-beds appear to be somewhat variable, so that there is no definite line of demarcation between the two formations.[1]

In both of these publications the Romney shales were limited below by the Monterey sandstone, which later has been shown to be the southern continuation of the Oriskany sandstone of New York,[2] as was stated many years earlier by Hall.[3] The upper limit of the Romney shales was the base of the Jennings formation, which has been shown to correspond with the Genesee shale, Portage formation (including the Sherburne, Ithaca, and Enfield members), and Chemung formation.[4]

While Chief of Division of Appalachian Geology of the Maryland Geological Survey the writer studied with some care the outcrops in the vicinity of Romney, Hampshire County, in northeastern West Virginia. Particular attention was given to the exposures of the Romney shales, while the subjacent and superjacent formations were also examined, and because Romney is the typical locality for this formation it is believed that a somewhat fuller account than has yet been published is desirable.

[1] *Geologic Atlas of the United States*, Folio 14, p. 2, col. 4.

[2] Schuchert, *Bulletin Geological Society of America*, XI (1900), 271, 312–15; Prosser, *Journal of Geology*, IX (1901), 416; Schuchert, *Proceedings United States National Museum*, XXVI (1903), 420; and Rowe, Schuchert, and Swartz, *Maryland Geological Survey, Lower Devonian*, text (1913), p. 90.

[3] Geological Survey of New York, *Palaeontology*, III (1859), 40, 401.

[4] Prosser, *Journal of Geology*, IX (1901), 419–21; Prosser, *Maryland Geological Survey, Middle and Upper Devonian*, text (1913), pp. 345–49; Swartz, *ibid.*, pp. 423–34.

DESCRIPTION OF WEST VIRGINIA SECTIONS

Cliffs near South Branch of the Potomac River.—The nearest outcrops of any considerable extent are the cliffs of argillaceous blue pencil shales by the side of the road southwest of Romney and near the bridge over the South Branch of the Potomac River. Most of these shales contain small iron-like concretions and on weathering are generally stained a brownish or iron color. There is a decided tendency in most of them to split into narrow, long, pencil-shaped pieces. The cliff in places is from 20 to 25 feet in height; the cut continues for some distance and affords a fine exposure of this part of the Romney shales. The dip is not great at this locality, and soon after crossing the river to the west the Oriskany sandstone is reached.

Fossils are rare, but a few occur in some of the layers. *Liorhynchus limitare* (Vanuxem), *Pterochaenia fragilis* (Hall), *Productella Ambocoelia*, a Crinoid stem, and perhaps a few other species occur. *Pterochaenia fragilis* (Hall) is known to occur in New York from the Marcellus shale to the Chemung formation inclusive, while *Liorhynchus limitare* (Vanuxem) is confined to the Marcellus shales and regarded as one of its most characteristic fossils. These bluish to blackish shales, containing the characteristic species—*Liorhynchus limitare* (Vanuxem)—of the Marcellus shale of New York, occur in at least the lower half of the Romney formation and are correlated with the Marcellus shale of New York.

Outcrops from Romney to Hanging Rock.—The valley road between Romney and Hanging Rock, which is down the river to the north of Romney, affords frequent opportunities to study the Romney formation, and the exposures may be considered as typical for its upper part.

Ledges of bluish-gray to grayish shale occur along the road by the side of Big Run, about one-half mile north of Romney, some of which has a decided concretionary tendency, forming large irregular layers. Fossils occur abundantly in certain of these layers, especially species of *Chonetes* and *Camarotoechia*, the latter on account of their abundance forming layers that are

partly calcareous. The following species were obtained at this locality:

Chonetes mucronatus Hall
Chonetes scitulus Hall
[The specimens are larger than the figures of this species, but in this respect do not differ from New York specimens in the Hamilton which are so referred.]
Chonetes lepidus Hall
Productella sp.
Camarotoechia congregata (?) (Conrad)
Camarotoechia sappho (?) Hall
Tropidoleptus carinatus (Conrad)
Cyrtina hamiltonensis Hall
Spirifer mucronatus (Conrad)
Spirifer granulosus (Conrad)
Ambocoelia umbonata (Conrad)
Vitulina pustulosa Hall
Grammysia sp.
Nucula corbuliformis (?) Hall
[External and internal impression.]
Nucula varicosa (?) Hall
Palaeoneilo constricta (Conrad)
Modiomorpha concentrica (Conrad)
Pleurotomaria (*Bembexia*) *sulcomarginata* Conrad
Macrochilus hamiltoniae Hall
Orthoceras sp.
Crinoid stems.

Dr. Kindle also stated that Hamilton fossils "may be collected near the railroad station at Romney."[1]

At the next rock-cut on the highway, about one mile north of Romney, the shales vary in color from bluish gray to gray and are rather more argillaceous than those of the preceding exposure. A number of small iron-like concretions occur in them, but fossils are not nearly so abundant as at the following locality and in some of these shales they are very rare. Toward the southern end of the bank they are more abundant and species of *Chonetes*, *Bellerophon*, *Orthoceras*, and some other forms occur. The complete list follows:

Chonetes mucronatus Hall
Chonetes scitulus Hall
Ambocoelia umbonata (Conrad)

[1] *United States Geological Survey, Bulletin 508*, p. 41.

Nucula bellistriata (Conrad)
Palaeoneilo constricta (Conrad)
Bellerophon sp.
Bellerophon (*Patellostium*) *patulus* (?) Hall
Loxonema hamiltoniae (?) Hall
Orthoceras sp.
Orthoceras constrictum (?) Vanuxem
Spyroceras crotalum (?) (Hall).

By the side of the highway about 2¼ miles north of Romney are outcrops of blue, arenaceous shale, the dip of which is very heavy, being nearly 80° east. These shales split into medium-sized pieces which do not disintegrate very rapidly into soil. The color is distinctly bluish and there are calcareous bands composed largely of fossils, *Spirifer mucronatus* (Conrad) being the most abundant species. Other Brachiopods are common, as *Tropidoleptus carinatus* (Conrad), *Chonetes*, etc., while there are also some small Lamellibranchs and Gastropods, but the species of these last two classes are rare. It is essentially a Brachiopod fauna of the Hamilton formation. The complete list of the species obtained at this locality is given below:

Stropheodonta (*Leptostrophia*) *perplana* (Conrad)
Chonetes mucronatus Hall
Chonetes scitulus (?) Hall
Chonetes setiger (Hall)
Chonetes lepidus Hall
Chonetes vicinus (Castelnau)
Productella sp.
Tropidoleptus carinatus (Conrad)
Cyrtina hamiltonensis Hall
Spirifer mucronatus (Conrad)
Vitulina pustulosa Hall
Grammysia lirata (?) Hall
Nucula randalli (?) Hall
Pterinea flabellum (?) (Conrad)
Aviculopecten sp.
Pleurotomaria (*Bembexia*) *sulcomarginata* Conrad
Loxonema hamiltoniae Hall
Tentaculites bellulus Hall
Orthoceras subulatum (?) Hall
Dalmanites (*Cryphaeus*) *boothi* (Green)
Crinoid segments.

This exposure is very similar to hundreds of outcrops in the arenaceous Brachiopod zones of the Hamilton shales in New York, and any geologist who has studied that formation in New York would at once recognize the great similarity in lithologic appearance. This, together with the presence of a Hamilton fauna, apparently justifies the writer in correlating that part of the Romney formation shown in the highway cuts north of Romney with the Hamilton formation of the standard New York section.

Hanging Rock.—About four miles north of Romney is a high cliff known as Hanging Rock, where the South Branch of the Potomac River has cut a deep and narrow gorge from the east to the west through Mill Creek Mountain. The river makes a big loop, and at the highway bridge south of Springfield it turns and cuts back through the ridge to its eastern side. At Hanging Rock, on the northern side of the river, is a high cliff in which the anticlinal arch of the fold in the Oriskany sandstone is finely shown. A fair idea of the appearance of this cliff may be gained from the halftone (Fig. 1.) At the western end of the gorge, by the side of the highway just after crossing the railroad track, is an excellent exposure of fine black to drab shales containing small concretions and some fossils. In one layer at least are numerous specimens of *Phacops cristata* Hall associated with *Dalmanella*, *Chonetes*, *Ambocoelia* and the representatives of a few other genera. The rocks are slightly arenaceous and break into small, somewhat elongated pieces on weathering. The layer in which the fossils occur abundantly is somewhat arenaceous and not very much above the top of the Oriskany sandstone. The species collected in this shale are given in the following list:

Stropheodonta sp.
Chonetes cf. *lepidus* Hall
[Two small imperfect internal impressions.]
Dalmanella lenticularis (Vanuxem)
Cyrtina hamiltonensis (?) Hall
[Small imperfect specimen.]
Ambocoelia umbonata (Conrad)
Styliolina fissurella (Hall)
[Abundant on some of the blocks.]
Phacops cristata Hall

In the above fauna, *Dalmanella lenticularis* (Vanuxem) is confined to the Onondaga formation; *Cyrtina hamiltonensis* Hall occurs in the Onondaga, Hamilton, and Portage; *Ambocoelia umbonata* (Conrad) from the Onondaga to the Chemung inclusive; *Styliolina fissurella* (Hall), southern Onondaga shale, Marcellus, Genesee, and Portage black shales; *Phacops cristata* Hall, Oriskany of Ontario, elsewhere the Onondaga. These species all

FIG. 1.—Hanging Rock, West Virginia, where the South Branch Potomac River has cut through Mill Creek Mountain. The rock cliff is Oriskany sandstone.

occur in the shales from Pennsylvania across Maryland and the Virginias which Dr. Kindle has correlated with the Onondaga of New York,[1] and since they came from the same stratigraphic horizon—the lower part of the Romney shale not far above the Oriskany sandstone—it appears that this part of the Romney ought to be correlated with the Onondaga. This correlation is in agreement with Dr. Kindle's statement that "the Onondaga shale

[1] *Journal of Geology*, XIX (1911), 97; *United States Geological Survey, Bulletin 508*, 1912; and *Maryland Geological Survey, Middle and Upper Devonian*, text (1913), pp. 48, 49, and 53–59.

member may be seen along the road through Mill Creek Mountain Gap, 1½ miles southwest of town [Romney]. Here the section shows an abrupt transition from the gray, highly calcareous coarse sandstone full of Oriskany fossils to drab shale approximately 100 feet thick, with some calcareous bands containing the Onondaga fauna. Fossils are not so readily found as at many other localities. Among those collected are *Anoplotheca acutiplicata*, *Dalmanella lenticularis*, and *Ambocoelia umbonata*."[1]

It is believed that the above account shows conclusively that the Romney shales in their typical locality near Romney, West Virginia, represent the southwestern continuation of the Onondaga limestone, Marcellus shale, and Hamilton formation of New York.

Outcrops south and north of Springfield.—North of Hanging Rock the valley road crosses a ridge of Oriskany sandstone and then for some distance south and north of Springfield the outcrops are mostly of thin black shales closely resembling lithologically the Marcellus shale member of the Romney formation. North of Springfield and perhaps three miles south of Green Spring in the cut of the Romney branch of the Baltimore and Ohio Railroad are black, argillaceous shales, and interstratified with the shales are several limestone layers, some of which are six inches in thickness. The limestone is very dark gray and blue to almost black in color, with a decided petroleum odor. It contains iron pyrites and is crossed by joints filled with calcite. There is a dip to the west and also to the south, and about 20 feet of rock is shown in the cut. Some of the layers of limestone contain a large number of small fossils, especially of the small species of *Ambocoelia*, which the writer has named *A. virginiana*.[2] Part of the limestone contains large numbers of this species, the best specimens of which came from this cut. When studied in the field these shales were referred to the Marcellus member; but it is possible that they occur lower in the Romney shale in that part of the formation which later has been correlated with the Onondaga.

Outcrops southeast of Romney.—To the east of Romney is the Jennings formation, succeeded by the Hampshire formation, the

[1] *United States Geological Survey, Bulletin 508*, p. 41.

[2] *Maryland Geological Survey, Middle and Upper Devonian*, text (1913), p. 202.

latter named from exposures in this county. There are only poor outcrops of the Jennings along the highway east of Romney, but farther east, in ascending the western slope of the northern part of South Branch Mountain, are excellent outcrops of red argillaceous shales belonging in the Hampshire formation. By the side of the highway, a short distance east of the first bridge over Big Run, are outcrops of sandstones and shales. The top and bottom layers consist of somewhat shattered, greenish to greenish-gray micaceous sandstone which breaks into blocks. Between the sandstones are thin, fissile, argillaceous shales, principally reddish-brown in color. The dip varies in different parts of the exposure from 11° to 14° S.E. No fossils were found, and the ledge is probably in the lower part of the Hampshire formation.

The rocks exposed by the side of the road east of this locality and up the valley of Big Run are mainly argillaceous shales, though there are some green ones, and interstratified with all of them are sandstones which are coarse-grained, greenish-gray in color, and massive. These outcrops occur along the highway below and in the vicinity of the locality known as the Peach Orchard, where, on top of the hill, the rocks are mainly red shale and the soil red in color, owing to their decomposition. These outcrops by the road and on the hill furnish a typical exposure of the Hampshire formation in northeastern West Virginia.

In Darton's account of the Devonian formations of central Virginia he says that "The Hampshire formation has yielded only a few plant remains which throw no light on the equivalency of the formation, but no doubt it comprised the representatives of the Catskill in their entirety or in greater part."[1] As stated by Darton, fossils are rare in this formation, as is the case in the corresponding one in Pennsylvania and New York; but the lithologic appearance and stratigraphic position agree, in general, with those of the Catskill formation, which has been shown to be a local one in New York, scarcely represented in the southwestern part of the state, while in the southeastern or Catskill Mountain region it has replaced all of the Chemung and the greater part of the Portage of western New York. It has been further shown that to the east of the

[1] *American Geologist*, X, 18.

Susquehannah River, in New York, the red rocks at first are interstratified with those which contain Chemung fossils, while farther east the red and greenish shales and sandstones replace all the rocks with the lithologic characters of the Chemung and its fauna disappears. Below the Chemung and in the midst of what corresponds to the Portage stage of western New York is another mass of red and greenish shales and sandstones called the Oneonta formation, which extends west to the Chenango Valley. In Delaware County the upper reds of the Oneonta and the lower of the Catskill unite and, in the Catskill Mountain region, extend downward into what is called the Sherburne sandstones, which represent the lower part of the Portage stage of western New York. In Pennsylvania, on following this mass of red rocks to the southwest it is found that they begin later, the change being gradual, and that the faunas of the Portage, including the Ithaca in Maryland and Chemung stages, reappear, as is the case in the southern part of western New York.[1]

The rocks exposed along the road following the highway toward Adams Mill, West Virginia, are largely red shales with some sandstones. Near the top of the hill are conglomerate layers, and across the upland the rocks are mostly red argillaceous shale, and all belong in the Hampshire formation.

In general it may be said that to the east of Romney the rocks belonging in the belt of the Jennings formation are mostly covered, while on the slope of the hill following the roads leading toward Frenchburg and Adams Mill are frequent outcrops of red shale interstratified with red sandstones belonging in the Hampshire formation. On top of these hills the red shales are most conspicuous, and decomposing rather readily into soil make good farming land.

Mill Creek southwest of Romney.—Mill Creek has cut a deep and narrow gorge through Mill Creek Mountain to the southwest of Romney which is followed by the highway from Romney to Moorefield Junction. In this gorge are high cliffs of the Oriskany sandstone which show excellently the massive nature of the formation.

[1] *Seventeenth Annual Report State Geologist* (New York), in which the writer has discussed this question.

The high cliff of Oriskany sandstone at the eastern end of the gorge is shown in Fig. 2. By the roadside not far west of the eastern end of Mill Creek Mountain Gap are layers of blue limestone interstratified with black argillaceous shales weathering to drab or brownish color. There are some fossils in the limestones, but they are infrequent. Along the road through this gap Dr. Kindle reported approximately 100 feet of drab shale with some calcareous bands succeeding the Oriskany sandstone, carrying an Onondaga

FIG. 2.—Eastern end of Mill Creek Mountain Gap, southwest of Romney, West Virginia. The steep cliff is composed of Oriskany sandstone.

fauna, and these basal shales of the Romney formation he correlated with the Onondaga of New York.[1]

Dr. Kindle further stated that: "The black shale representing the Marcellus is well exposed on the opposite side of the South Branch of Potomac River and appears to be barren. A little higher up the hill to the eastward on the west side of Romney the black fissile shale is succeeded by olive to gray slightly sandy shale with Hamilton fossils."[2]

[1] *United States Geological Survey, Bulletin 508*, p. 41.

[2] *Loc. cit.*

On the Mill Creek Road, $5\frac{3}{4}$ miles southwest of Romney or $1\frac{1}{4}$ miles east of Moorefield Junction, on the farm of Mr. Parker, is an outcrop of argillaceous shale. The dip is 35° N., 30° W. at this ledge, but in a short distance it changes. The shales are in general smooth, olive in tint, and in some that are a little mealy in texture are fossils, the most common species being *Pterochaenia fragilis* (Hall), *Buchiola retrostriata* (?) (v. Buch), *Goniatites*, *Coleolus*, etc.

This exposure is nearly all shale, and it belongs in the lower part of the Jennings formation. In fauna, lithologic appearance, and stratigraphic position it agrees closely with the Portage stage of the Upper Devonian of New York. It will be remembered that Dana united the Genesee, Portage, and Chemung to form his Chemung period, which corresponds very nearly, if not quite, to the Jennings formation.

On the bank of a run about one mile east of Moorefield Junction are olive, argillaceous shales by the side of the road, which alternate with thin, micaceous sandstones. No fossils were found, and the rocks are referred to the Portage stage. A dip of 45° S., 40° E. was noted. A ledge by the side of the road on the southern bank of Mill Creek, at the crossing directly east of Moorefield Junction is composed mostly of smooth, argillaceous, olive shales interstratified with thin sandstone layers which are rather micaceous. At the creek level is a sandstone layer two or more feet in thickness. The dip is about 2° west. No fossils were found and the rocks are referred to the Portage stage.

To the west of Moorefield Junction smooth, argillaceous, olive shales, alternating with layers of sandstone, occur along the highway and dip to the east. No fossils were found. These shales, as in the case of those east of Moorefield Junction, are considered as of Portage age and occur in the lower part of the Jennings formation.

Patterson Creek.—The next valley to the west of that of South Branch is that of Patterson Creek in Mineral County, and since in the volumes on the paleontology of New York certain Hamilton species are mentioned as occurring at Patterson's Creek, West Virginia, the outcrops in the lower part of the valley of this creek were examined. The Romney area of this valley is a continuation of the narrow band which lies between Collier and Nicholas moun-

tains, to the north of the Potomac River, in Allegany County, Maryland.

The railroad station and post-office now known as Patterson or Patterson Depot is the locality formerly known as Patterson Creek. There are no exposures in the immediate vicinity of the station, and the Baltimore and Ohio railroad track, along which are several cuts, was followed to the eastward. The first one east of the station shows massive Oriskany sandstone, the greater part of which is hard, light gray, and quartzitic, certain layers containing a considerable number of fossils, particularly specimens of *Spirifer arenosus* (Conrad), some of which are very perfect. At the eastern end of the cut the rocks dip sharply to the east, while farther to the west they are some distance above the track and nearly horizontal. In the second railroad cut east of the station is another outcrop of the Oriskany sandstone, which is also dipping to the eastward. In some of the partly decomposed layers of brown sandstone there are good specimens of *Spirifer murchisoni* Castelnau, associated with other Oriskany fossils. From near the center of the cut to the western end there is some of the very hard, quartzitic sandstone. The dip is about 23° S., 70° E.

At the first hill east of the cut described above are thin, bluish, arenaceous shales which form a low ridge. These shales are rather firm, and when crumbled the pieces are larger than those on the Williams Road to the north in Maryland. No fossils were found. These shales are apparently in the lower, fairly barren portion of the Hamilton member of the Romney formation. It is to be remembered in this connection that the lower part of the Hamilton formation consists of rather arenaceous shales, in which fossils are of infrequent occurrence. This area is the southwestern continuation of the belt of the Romney formation to the east of Collier Mountain in Allegany County, Maryland.

Along the hill by the highway about one mile south of Patterson Depot are outcrops of thin, bluish-black shales which are fairly argillaceous and split into thin pieces weathering to a brownish color. In certain layers fossils are rather thick, especially *Ambocoelia umbonata* (Conrad) and *Liorhynchus limitare* (Vanuxem). Specimens of *Coleolus* and small-winged Lamellibranchs were also

seen. On the ridge to the west is the Oriskany sandstone, dipping to the east, as shown in Rocky River and on the highway from the west perhaps one-sixth of a mile north of this locality. These shales contain the diagnostic species—*Liorhynchus limitare* (Vanuxem)—of the Marcellus shale and agree fairly well lithologically, as far as exposed, with those referred to the Marcellus shale in the Baltimore and Ohio railroad cut at the 21st Bridge, Maryland; consequently, they are referred to the Marcellus member of the Romney formation.

The valley to the east of this ridge along Patterson Creek is presumably composed of the Romney shale, which is probably in a small syncline. The steep wooded ridge to the east of Patterson Creek is the one in which occurs the first railroad cut east of Patterson Depot. The Patterson Creek Road was followed fully two miles from the station, and after the first exposure of the black Marcellus shales they are seen at frequent intervals at the roadside. The more southern shales examined, however, were rather coarser, somewhat more arenaceous, and less fossiliferous. The greater part of the Marcellus is slightly arenaceous and with less clay than is found in the exposures on the Williams Road to the northwest in Maryland, and the shales do not disintegrate so readily.

On Plum Run, two miles above Patterson Depot and below Mr. Robinson's house, are coarse, unfossiliferous, arenaceous shales varying to fairly thin-bedded sandstones of bluish-gray color. The zone, however, is clearly above the fine black shales of the Marcellus, which contain numerous specimens of *Liorhynchus limitare* (Vanuxem).

On Mr. Robinson's farm is a small stream to the southwest of Plum Run, on which are bluish shales and sandstones, some of the latter being very hard and slightly calcareous. It is probable that this sandstone zone corresponds with the lower one of the Hamilton member of the Romney formation on the Williams Road, Maryland. Specimens of *Spirifer mucronatus* (Conrad) and *Ambocoelia umbonata* (Conrad) occur sparingly. On the small hill to the south of the run are yellowish to greenish, very argillaceous shales, containing some fossils. *Chonetes* is abundant in certain thin layers while other species, occurring more sparingly, are *Spirifer mucro-*

natus (Conrad), *Spirifer audaculus* (Conrad), *Nuculites oblongatus* Conrad, *Phacops*, *Pleurotomaria ityx* (?) Hall, *Orthoceras telamon* (?) Hall, and *Macrochilus hamiltoniae* Hall.

In color and composition these shales are fairly similar to those above the lower sandstone on the Williams Road, Maryland; but on weathering they seem to have hardened to some extent. This shale contains, here and there, numerous small hard concretions as well as those of clay-iron stone, and it is also blotched with spots and streaks of dark-red color from the weathering of the iron. It is very probable that the specimens of *Nuculites oblongatus* Conrad, *N. triqueter* Conrad, *Palaeoneilo constricta* (Conrad), *P. tenuistriata* Hall, *P. virginiana* Hall, *Grammysia alveata* (Conrad), *G. arcuata* (Conrad) listed by Hall from Patterson's Creek, Va.,[1] and perhaps other species, came from this portion of the Romney formation. These outcrops have a lithologic similarity to beds across the Potomac River in Maryland which have been correlated with the Hamilton and are their southern continuation. Furthermore, these rocks contain fossils that are common in typical outcrops of the New York Hamilton, and for these reasons they are referred to the Hamilton member of the Romney formation.

NOTE ON THE CORRELATION OF THE MARYLAND HELDERBERG FORMATION

Dr. Swartz in his "Introductory" account of "The Helderberg Formation" states that "Rowe divided the Helderberg into the equivalents of the Manlius, Coeymans, New Scotland, and Becraft formations of New York. O'Harra, who prepared a report upon the geology of Allegany County about the same time, assigned the same limits to it."[2] Mr. Richard B. Rowe was one of the writer's students when he was professor of geology in Union College, where Mr. Rowe studied under his direction the Helderberg formations in their typical region in the Helderberg Mountains of eastern New York. Later Mr. Rowe went as a graduate student to Johns Hopkins University and began the field study of the Paleozoic formations of western Maryland during the summer of 1897. In

[1] *Palaeontology*, Vol. V, Part I, Lamellibranchiata II, 1885.

[2] *Maryland Geological Survey, Lower Devonian*, text (1913), p. 97.

1898 the writer became Chief of Division of Appalachian Geology of the Maryland Geological Survey, and with Mr. Rowe as his principal assistant devoted that summer to field work on the Paleozoic formations of western Maryland. We worked together in the field and were fully agreed concerning the correlation of certain limestones with the Coeymans, New Scotland, and Becraft formations of New York, which is believed to be the first definite recognition of them in Maryland. A statement of the relationship of this work was published by the writer in 1901,[1] together with the correlation of the Helderberg limestone with the New York formations.[2] This was recognized by Mr. George W. Stose, who wrote as follows:

> The presence in this region [Pawpaw-Hancock Folio] of a representative of the Helderberg group of New York has long been known, but in earlier reports of the United States Geological Survey and of the Maryland Survey it was not differentiated as a separate formation but was included with the underlying calcareous formations under the name Lewistown. In 1901, Prosser, reporting the results of studies made by him and his associates for the Maryland Geological Survey, later embodied in the State report [Vol. VI (1906), 133, 134], recognized in the Maryland rocks the various faunal divisions of the Helderberg of New York.[3]

The writer in his *Historical Review and Bibliography* of the Maryland Devonian summarized the contents of his 1901 article on "The Paleozoic Formations of Allegany County, Maryland."[4] The notice of "The Geology of Allegany County" in the same review contains the statement that "The nomenclature and classification of the formations are those adopted by Messrs. Clark, Prosser, and Rowe for the Maryland formations."[5]

[1] *Journal of Geology*, IX, (1901) 409.

[2] *Ibid.*, 415, 416.

[3] *Geologic Atlas of the United States*, Folio No. 179 (1912), p. 9, col. 1.

[4] *Maryland Geological Survey, Lower Devonian*, text (1913), p. 51.

[5] *Loc. cit.*

THE STRENGTH OF THE EARTH'S CRUST

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PART VII. VARIATIONS OF STRENGTH WITH DEPTH AS SHOWN BY THE NATURE OF DEPARTURES FROM ISOSTASY

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GEODETIC EVIDENCE AS TO LIMITING HEIGHTS AND WAVE-LENGTHS

Measurements of strength by maximum loads.—In the course of geologic time the internal forces of igneous intrusion and tangential compression, the external forces of erosion and sedimentation, have tended to strain the crust to its limits of strength, and the degree of isostasy which exists constitutes a measure of those limits. Small loads of large wave-length and large loads of small wave-length will tend to rise to maxima which may be used in connection with the theory of distribution of stress, as considered in Part VII, Section A, to give an approximate idea of the limits and distribution of strength in the crust.

The problem is to find the maximum vertical load and its relation to wave-length acting over areas which may be regarded as

forming roughly a harmonic series. The theory applies best where elongated unit areas are flanked by areas of opposite sign. But even where a single axis of uncompensated elevation or depression is surrounded with a region of mean elevation it may be regarded as a half of a wave sustained by the rigidity of the earth. The stresses in the vertical plane through its crest line would appear to be less than, but not so greatly different from, what they would be for a continued series. Where the load is of oval instead of zonal distribution the stresses would also be somewhat diminished in case the oval is surrounded by a neutral region, but if a series of ovals of opposite signs is analogous to two intersecting wave-series, although the stresses would be complicated and are not in general the sum of the stresses due to the separate series, yet it does not appear that the extreme maxima would be necessarily less than the sum, and many of the maxima would be greater than the maxima of the separate series.

Harmonic loads with short wave-lengths.—Horizontal compression builds mountain folds of which the individual ranges are clearly the results of compression and not of isostatic elevation. Erosion dissects an elevated country on a pattern of a certain scale, deep valleys of erosion separating the crust remnants above base-level which are not yet consumed. These actions produce a surface relief which corresponds roughly to various harmonic series of appreciable amplitude and wave-length, but in this section will be considered only the geodetic evidence of variable mass not isostatically compensated, much of the variation being due to the concealed heterogeneities of density.

The distance between Washington, D.C., and Hoboken, New Jersey, as estimated in Part V, Section B, is 326 km. Within this distance are three intermediate stations and the two limiting stations, each station showing a gravity anomaly opposite in sign to that of the adjacent stations. There must be then at least two wave-lengths. The average change of anomaly between the adjacent stations is 0.021 dyne. As it is wholly improbable that the stations are located at the crest lines of the waves, the whole amplitude may safely be taken as at least 0.026 and the wave-length 160 km.

Seattle and Olympia are 80 km. apart and the difference of their anomalies is 0.126 dyne. If the anomalies had been measured at the points giving maximum values they would certainly give a difference of at least 0.130 or perhaps 0.140, about five times the amplitude of the variations between Washington and Hoboken. This large value seems, however, to be exceptional for the United States and may constitute but a single wave. We may take it, however, as showing that the crust can sustain a harmonic load of 160 km. (100 miles) wave-length and total amplitude of 0.120 dyne. For the reasons explained in Part IV, especially on p. 304, the divisor to be used to turn this anomaly into the equivalence of rock measured in feet could not be over 0.0018 and a better figure for the interpretation of this short wave-length is 0.0015. This gives an amplitude of 8,000 feet (2,440 m.). The part below the mean level is then 50 miles wide and 4,000 feet deep, the adjacent positive parts of the wave being of equal dimensions but in the opposite direction. This is of the same order of relief as the larger ranges of folded mountains and intermontane valleys. The stresses which this harmonic series imposes upon the crust are shown by curve A of Fig. 18.

Maximum loads for mean wave-lengths.—In Part II it was argued that the evidence of anomalies from mountain stations showed regional compensation on the average probably to the outer limit of zone O, radius 166.7 km., diameter consequently 333.4 km.

Over the United States in general the intercepts of the areas of grouped deflections averaged 180 miles. The average diameter would therefore doubtless be as great as 200 miles (320 km.).

On Bowie's map of the New Method gravity anomalies,[1] it is seen that the distances from pronounced maxima to pronounced minima average 250–350 km.

From these several lines of evidence we may conclude with some confidence that the half wave-length for pronounced anomalies in the United States averages near 300–350 km. The wave-length is therefore 600–700 km. (373–435 miles). A wave-length of 600 km. will be taken. The pronounced maxima for these waves runs from plus or minus 0.030 to 0.060 dyne. The real maxima

[1] This article, Part II, *Jour. Geol.*, XXII (1914), 153.

are to be regarded in most cases as situated to one side of the stations and somewhat greater. But exceptional and local maxima must be smoothed out to form a part of the harmonic curve. It is, furthermore, the difference of anomaly between adjacent opposite phases which is the significant feature. This difference runs from 0.060 to 0.080. The latter figure will be chosen. For this wave-length, representing a certain unit area of attraction, the best divisor is perhaps to take 0.0024 dyne of anomaly as equivalent to 100 feet of rock. An anomaly of 0.080 dyne is on that basis equivalent to 3,330 feet (1,015 m.) of rock. The crust of the United States sustains, therefore, harmonic loads 600 km. (373 miles) in wave-length and 1,015 m. (3,330 feet) in total amplitude. The stresses which this wave-series imposes on the crust are shown by curve B, Fig. 18.

Departures from isostasy of large wave-lengths.—For the continent as a whole and in its relations to the ocean basins isostasy is nearly perfect; but the question rises here, how nearly? The first term of the gravity formula for the Vienna system of gravity observations is 978.046 dynes. The first term for the Potsdam system is 978.030. The first term for the United States system after rejecting the Seattle anomalies is, as shown by Bowie, 978.038 dynes. These respective systems differ as a whole by these amounts. We have no right to assume that any one is absolutely correct. The whole of the United States system may lie a little above or below the level giving isostatic compensation with respect to the average surrounding ocean basins, or with respect to the entire earth. The mean value for the United States suggests, however, that, as a whole, the continent lies within a few hundred feet, possibly less than one hundred feet, of the level which would give perfect isostatic equilibrium.

Let us consider next its larger parts. These can be compared with each other and with the United States as a whole. Although, as discussed in Part IV, the map of gravity anomalies lacks detail, the grouping of many stations of like sign into large areas gives confidence in the conclusion that there are regional departures from isostasy. These are of two or three orders of magnitude, of which the areally smaller have been discussed. To bring out the

areally larger we must draw boundaries about large regions which show a dominance of anomalies of one sign. These boundaries, however, must be taken so as to give compact unit areas, so as not to obtain an unreal result by the political expedient of gerrymandering the districts.

Select as a center the point whose geographic co-ordinates are lat. 42°, long. 102°. Describe about this center a circle of 850 km. radius. This includes an area equal to 29 per cent of the area of the United States. It should be taken as including the negative anomaly station 99 on its southern border. This circle covers a large positive region which could be made still more positive by an extension of its boundaries to the northeast over Wisconsin and Michigan. Within this circle are distributed with a fair degree of uniformity 31 of the 122 gravity stations of the United States. The mean with regard to sign of the anomalies of these 31 stations referred to the United States mean with regard to sign is +0.010 dyne. As the mean without regard to sign of all stations in the United States excluding Seattle is only 0.018, it is seen that this positive region stands out clearly from the general average.

West of this circle and, on the south, to the west of long. 107° there are 21 stations, including one of the two Seattle stations. These mark a broad region of negative anomaly. The mean anomaly with regard to sign is −0.017 dyne. There seems to be no reason for completely omitting the exceptionally large Seattle anomalies. One of them has therefore been retained, but if both are omitted the mean is still −0.013. The value of −0.017 will here be adopted. The difference of the means of the central and western regions is consequently 0.027 dyne. Let these be regarded as the positive and negative phases of an harmonic wave and the mean departure of the two phases becomes 0.0135 from each side of the mean plane. Now it may be computed for a harmonic wave represented by the formula $y=A \sin Bx$ that the mean height of the wave above the mid-plane is 64 per cent of the crest height. From mid-plane to crest of this wave-series is therefore 0.021. From the large negative anomalies along the Pacific coast it would seem that this negative zone must extend somewhat further. The wave-length of this series is consequently between

2,600 and 3,000 km. A mean value of 2,800 km. (1,740 miles) will be chosen. From the breadth of half a wave-length it appears that 0.0034 dyne of anomaly may be taken as equivalent to 100 feet of rock. This gives the crest and trough as 625 feet (190 m.) from the mean plane, a total amplitude of 1,250 feet (380 m.). The stress-differences which this wave-series throws upon an earth elastically competent throughout to bear the stresses are shown by curve C, Fig. 18. Helmert has published an extensive paper dealing with the force of gravity and the distribution of mass in the crust of the earth,[1] to which the writer's attention has been called recently by Professor Pierpont, of the mathematical department of Yale University. In this paper Helmert adopts the hypothesis of regional isostasy and finds his results confirmatory of it, but not in accord with the hypothesis of close and local isostatic adjustment. His work is especially valuable as confirmatory of the present conclusions, since it deals with regions outside of the United States. As he does not, however, compute the corrections due to the distant large irregularities of topography, his figures cannot be directly compared with Hayford's New Method anomalies. Nevertheless his conclusions as to the existence of broad regional excesses or defects of mass are comparable to those here reached. Under the section on the horizontal displacement of compensation and extended excesses and defects of mass[2] he sums up part of the evidence in the following statement: "We have then to deal with a continuous region of positive total gravity disturbance in Europe 1,000 km. broad and also with a region of negative disturbance in Asia of at least 500 km. breadth, both possessing great linear extension."

RELATIONS OF ACTUAL STRESSES TO THE SUM OF HARMONIC WAVES

Both Darwin and Love point out that the actual stress-differences imposed by the superposition of different harmonic waves is not in general the sum of the individual stress-differences. Darwin, however, states the special conditions under which the

[1] "Die Schwerkraft und die Massenverteilung der Erde," *Ency. Math. Wissenschaft*, Band VI, 1, B, Heft 2 (1910), pp. 85–177.

[2] *Op. cit.*, pp. 152–54.

resultant is the sum of the individual stress-differences.[1] The three waves which have been considered are types which coexist and are superimposed. The total stress which they give would vary from

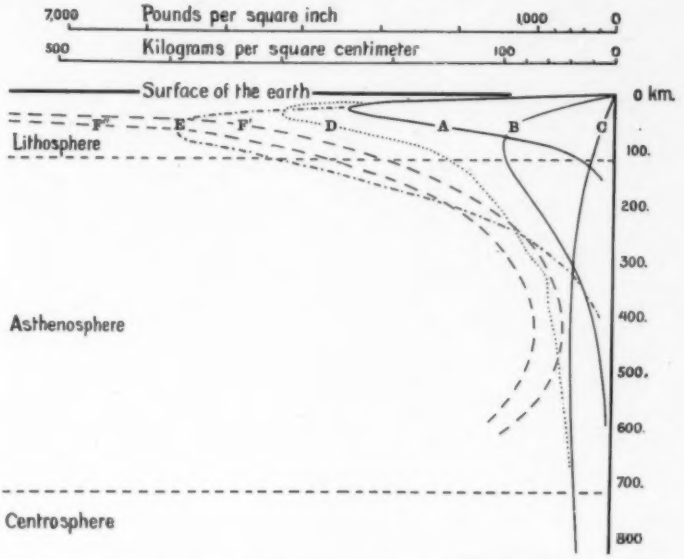

FIG. 18.—Stress-curves for harmonic waves on an earth elastically competent throughout, the waves representing departures from isostasy in the United States as given by analyzing the geodetic data into the following harmonic waves:

A, wave-length 160 km., amplitude 0.120 dyne = 2,440 m. of rock.

B, wave-length 600 km., amplitude 0.080 dyne = 1,015 m., of rock.

C, wave-length 2,800 km., amplitude 0.042 dyne = 380 m. of rock.

D is the sum of A, B, and C.

E, wave-length 400 km., amplitude 0.366 dyne = 4,000 m. of rock (from the Pacific Ocean).

F′, curve of strength suggested by geodetic evidence from the United States.

F″, curve of strength suggested by geologic evidence from various parts of the world.

complete neutralization up to their sum as a possible maximum. Curve D represents such an addition of A, B, and C, Fig. 18. There are reasons why this curve may be taken as a fairer representation of the maximum stress conditions under the United States than

[1] *Scientific Papers*, II, 492.

any individual curve, even though there may be no place where the culminating phases of like sign all coincide and become additive. These reasons are found in the subsurface location of those loads due to outstanding density and also to the added stresses due to isostatic compensation. These causes result in throwing a greater stress upon the outer parts of the lithosphere and also serve to broaden downward the stress diagrams. Furthermore, the stresses due to isostatic compensation of continents would appear to be in reality much greater under the margins than the small values computed by Love,[1] since he has taken the continents as having the broad sweeping surfaces of a harmonic nature, whereas, as a matter of fact, the continents slope off rather abruptly to the depths of the oceans. Facing the Pacific in fact, the two Americas show high mountain elevations. This would cause the stresses in the vicinity of these continental margins to resemble those imposed by a great mountain chain and its isostatic compensation rather than those imposed by the breadth of a continent. If isostatic compensation is complete under mountain slopes, Love shows for the cases computed by him that the maximum stress is about equal to that given by a column of rock one-fourth the total height from mountain crest to valley bottom. If the abyssal slopes of the continental platforms be taken as averaging 3–4 km. in elevation and 50–100 km. in width, it is seen that, even if fully compensated, they add stresses to the crust which may approach in magnitude one-half of the stresses shown by curve A of Fig. 18. The extreme depths of slope are much greater and it is clear that isostatic compensation cannot be exact under these great reliefs. Therefore we may conclude that curve D does not overestimate the maximum stresses imposed by the irregularities of the crust, both compensated and uncompensated, as indicated by geodetic evidence within the United States and especially along its ocean borders. This investigation, however, has been of a general nature and is designed merely to establish an order of magnitude. It remains for future work to make more precise analyses for each locality from the data which may be acquired, and especially to investigate quantitatively the problems offered by critical areas.

[1] Some Problems of Geodynamics (1911), chap. ii.

GEOLOGIC SUGGESTIONS AS TO MAGNITUDES OF CRUSTAL STRESSES

Submarine geanticlines and geosynclines.—Here will be considered some geologic illustrations of departures from isostasy, arranged in order of harmonic wave-length. They are to be compared with the results obtained from the study of the geodetic data. Most of the geologic evidence is merely suggestive, not conclusive, since diametrically opposite opinions are held as to the probability of the visible load being offset by an invisible compensation. As suggestions, however, they are none the less valuable, and point the way to needed geodetic observations.

The mountain folds advance from Asia over the floor of the Pacific Ocean, forming the system named by Suess the Oceanides. Mostly hidden beneath the ocean surface, they have been but little affected by erosion. Their ridges and deeps mark the greatest mountain reliefs of the globe. It is probable that here, if anywhere, tangential pressures have forced the crust into folds whose height combined with span is as great as the strength of the crust can endure. To what degree the elevations and depressions are compensated is, however, unknown, and the great arches are supported in part by the lateral pressure of the ocean water. It is quite possible if not probable that appreciable changes of deep-seated density may accompany the growth of such ridges, especially as they mostly exhibit a volcanic activity and are to a greater or less extent structures built up by igneous extrusion. It is not at all probable, however, that they are completely or possibly even largely compensated, but where the mountain folds and trough-like deeps broaden into plateaus or anti-plateaus the presumption is strengthened that the forms may there be isostatically compensated to a large degree. Such plateaus or anti-plateaus cannot then be used in the present argument. The ridges and troughs, however, show in their forms, as has been stated, modes of construction which are not conditioned on isostasy. Let attention be turned then to the folds of the ocean floor.

Passing from west to east, first may be noted between the Philippines, Borneo, and New Guinea a complex of ridges and basin-like deeps. The larger wave-length of that region runs from 300 to 500 km. The Ladrone Islands and Nero Deep give a

distance of about 150 km. from crest to trough and a wave-length of 400–500 km., these folds and many others exhibiting a lack of symmetry. The existence of strong folding pressures and a tendency to overthrusting and secondary vertical faulting seem to be expressed therein. The great fold passing north from New Zealand and showing as the Kermadec and Tonga islands with their fore-deeps gives a distance from crest to trough of 120–180 km., a wave-length of 400–500 km. Lower California and the troughs on each side show a wave-length of 350 km. The region of the Lesser Antilles is tectonically a northward branching of the Andean mountain system and shows, like the folds of the Pacific, crustal undulations with a wave-length of 350 km. We may conclude then that these folds of the ocean floor have a marked tendency to a wave-length of 300–500 km., there being commonly one great asymmetric fold passing out into subordinate marginal folds. The volcanic chain of the Hawaiian Islands shows, however, no related deeps and has a half wave-length of about 200 km.

Hayford and Bowie have given the New Method anomalies for a few stations in these regions.[1] A portion of the data has been abstracted and given in Part IV of the present article.[2] Four observations of Hecker's for the Tonga Plateau and Tonga Deep are given. They may not be of high value, since the method has been criticized as not possessing accuracy comparable to observations made by pendulum upon land. Furthermore, the four observations, two over the plateau and two over the deep, are spread through a distance of 5,100 km. along the axis of the structure instead of being taken on a transverse section. Nevertheless, as the reliefs and the corresponding anomalies are all of great magnitude, the errors become relatively small and the mean of the observations is therefore of some value. The two New Method anomalies for the Tonga Plateau give a mean of +0.202 dyne, the depth of water being 2,700 m. The two New Method anomalies for the Tonga Deep give a mean of −0.172 dyne, the depth of water averaging 7,500 m. If the amplitude

[1] *The Effect of Topography and Isostatic Compensation upon the Intensity of Gravity*, 1912, p. 81.

[2] *Jour. Geol.*, XXII (1914), 311.

of the uncompensated portion of the crust-waves be measured in terms of anomaly by taking the algebraic sum of the anomalies over the plateau and the deep, a total amplitude is obtained of 0.374 or a half-amplitude of 0.187. On the island of Hawaii an observation on Mauna Kea at an elevation of 3,981 m. gave a New Method anomaly of +0.183, almost the same figure as the half-amplitude for the great Tonga crust-wave.

Helmert has discussed the gravity disturbances found in the Hawaiian Islands and states of them: "For the Hawaiian Islands it must be concluded on the whole that a part of the mass gives rise to positive gravity disturbances and only the remainder is isostatically supported. If the disturbances were produced solely by the mass of the islands the values of Δg and $\Delta g''$ [the disturbances of gravity] would be somewhat greater than they are found."[1]

From this review of the mountain chains of the Pacific it may be concluded that the ocean floor can sustain a harmonic wave-length of 400 km. which gives an anomaly at the crest lines as great as that observed on Mauna Kea, 0.183 dyne. To interpret this as an equivalent load of rock a divisor must be selected. The divisor depends upon the depth and distribution of compensation and the area of the region of outstanding mass. As shown in Part IV, p. 311, 0.0024 dyne might reasonably be chosen as the amount of anomaly equivalent to 100 feet of rock, but for these great loads it is desirable to lean toward the side of an underestimate. Therefore 0.0030 will be taken as the divisor. This gives 1,868 m. as the crest height of the uncompensated portion of the Hawaiian mountain chain. The same applies to the Tonga fold. If, however, 0.0024 should be chosen as the divisor, then 0.183 dyne of anomaly would correspond to a half-amplitude of 2,334 m.

It may be taken then as fairly certain that these great mountain chains show reliefs which depart as much as 2,000 m. above and below the mean level which would give perfect isostasy. It may be concluded in consequence that the oceanic crust can sustain a harmonic wave-length of 400 km. with an uncompensated amplitude measured by 4,000 m. of rock. The diagram of

[1] "Die Schwerkraft und die Massenverteilung der Erde," *Ency. math. Wissenschaften*, Band VI, 1, B, Heft 2, (1910), p. 133.

stress-differences for this is shown in curve E, Fig. 18. But if the rock has a density of 2.67 and the sea-water a density of 1.03, this corresponds to an amplitude beneath the ocean surface of 6,513 m. of uncompensated rock. This is only about two-thirds of the maximum relief which is observed, so that it is well within the limits of possibility. These few available figures suggest that the sharp submarine ridges and deeps may not be more than one-third or two-thirds compensated.

The Niger delta.—Reverting to the discussion of the Niger delta given in Part I, it is seen that there is no evidence of depression around its margin. It may be taken then as the positive half of a harmonic wave well within the limits of crustal strength. If the section of the delta be taken as given in Figs. 3 and 4, pp. 31, 43, it is seen that the load is disk-like in form, instead of being indefinitely elongated at right angles to the section in accordance with the form of a zonal harmonic. It seems likely, because of these two departures from the nature of a harmonic series, that the stresses beneath it are not more than half of those which would be given by the completed harmonic curve. As it is merely the order of magnitude of the stress-differences which we may hope to attain we may proceed in accordance with these rough assumptions. It is seen that the section of the delta shows a half wave-length of about 300 km. and a maximum thickness equivalent to 1,650 m. of rock upon land. This corresponds to the half-amplitude or thickness above the mean plane. If half of this is taken as a measure of the stress, it gives a wave-length of 600 km. and a total amplitude of 1,650 m. The stress-curve for this harmonic series is 60 per cent larger than the stresses due to the outstanding masses of the same wave-length as given by the geodetic evidence in the United States and shown in curve B, Fig. 18. As the estimate from the Niger delta is very imperfect and unchecked by pendulum observations reduced by the New Method, the stress-curve is not plotted.

The existing continental ice sheets.—Two ice sheets of subcontinental proportions remain in existence, the Greenland and Antarctic. They form great plateaus sloping upward from the margins; the Greenland sheet reaching elevations at its center

between 9,000 and 10,000 feet, the Antarctic attaining to about 11,000 feet. The average thickness of the ice must be thousands of feet in each case. The development of these ice caps during the refrigeration of climate which marked the later Tertiary must have imposed upon the crust great loads of wide span. If isostatic equilibrium was previously complete to a large degree, the ice mantles should give valuable measures of crustal strength. For this purpose, however, a set of gravity measurements should be carried inland and reduced by the Hayfordian method. The facts that these two ice-mantled areas are both high plateaus, and that no other adjacent unglaciated land is of similar topographic character, suggest that these regions may be competent to carry great thicknesses of ice without isostatic yielding. There is no present basis, however, for making a quantitative estimate. It must be borne in mind, furthermore, that the ice mantle is only about one-third of the density of rock and that lofty mountains exist in both regions, showing that these lands would possess considerable mean elevation even without the presence of the ice. The effect of the difference of density between ice and rock may be appreciated by considering that an ice sheet 3,000 feet in thickness would possess the same mass as a layer of rock 1,000 feet thick. For isostasy to remain perfect after the development of this ice sheet, the crust would have to sink 1,000 feet, but the surface of the ice would still be 2,000 feet above the former level and give an appearance of load which would not in reality exist.

This is a problem meriting research for several reasons. A knowledge of the load which is sustained by these regions would show to what degree the warpings connected with the extinct Pleistocene ice sheets were mere elastic responses to load, to what degree they marked subcrustal plastic flow working toward isostasy. The results could be applied also to the problem as to how far from isostatic equilibrium a continent might come to lie as a result of continent-wide base-leveling in a period of geologic quiet. It seems not impossible that the stress-curve due to the portion of the glacial load which is elastically sustained would give stress-differences greater at a depth of 300–500 km. than those shown by curve C, Fig. 18. Such an investigation may then

be an essential factor in measuring the maximum strength of the lithosphere and more especially the asthenosphere.

Accordance of geologic with geodetic evidence.—The United States and its bordering ocean bottoms is a region of moderate reliefs as compared to the great folds of the ocean floor or of the continent of Eurasia. The geologic forces of folding and uplift have not worked here with their greatest intensity and the central and eastern half of the continent has been affected by the world-involving Cenozoic diastrophism to only a moderate degree. It is to be expected then that the greatest strains upon the crust, the maximum departures from isostasy, would not be found here. In accordance with this expectation it has been seen that by far the greatest New Method gravity anomalies are found in other regions and associated in most cases with the greater reliefs of the globe. The geologic evidence is in harmony; the amount of uncompensated relief, parallel to the geodetic evidence, is greatest for the lesser wave-lengths; but, throughout, the geologic evidence suggests that the actual burdens which can be borne by the crust, as found in regions of culminating stress, are appreciably greater than those detected by geodetic methods as existing in the region of the United States.

If, in some past ages, as during the Appalachian or Sierran revolutions, strains were generated in this continent as great as those found now in some other regions, it would appear that the slow changes of geologic time, of erosion and crustal readjustment, have partially eased the crust of its load. We may have, then, a variable crustal strength—a maximum strength exhibited during and following the crises of great diastrophism; another, lesser strength, which measures the loads which the crust without failing can bear through all of geologic time.

ADJUSTMENT OF LOADS TO THE DISTRIBUTION OF STRENGTH

It has been seen that the departures from flotational equilibrium may become very notable and are of greatest vertical magnitude for wave-lengths from 100 up to 400 km. The strains generated by these loads, if distributed through an elastic crust, consequently

reach maximum values at depths not exceeding 64 km. Is this because the earth shell below the zone of compensation is strong, but for some unrelated reason free from large stress-differences, or is there an absence of such stress-differences because this shell is too weak to bear them? If the latter is true, then the relations of amplitude to wave-length which have been developed in this chapter offer additional proofs of the reality of the existence of the asthenosphere.

The geologic evidence on the evolution of continental structures and elevations leads to the conclusion that the distribution of stress-differences must be in reality the result of the existence of a zone which cannot carry large distortional strains, as may be seen upon brief consideration.

The internal activities of igneous intrusion and of tangential compression do not in themselves work toward isostatic equilibrium, but merely toward accentuation of relief. Erosion and sedimentation, while tending to destroy this relief, are not agents tending to create, but to destroy, such isostatic relations as have developed. All of these activities work on a continental or subcontinental as well as on an orogenic scale, as seen in the Cenozoic history of the broad Cordilleran province, yet while the orogenic departures from isostasy are vertically very great, the continental departures are very moderate. For the latter there must be then some more narrowly limiting condition. This corresponds to the incapacity of a deep zone, the asthenosphere, to carry large stress-differences and the incapacity of the lithosphere in spite of its greater strength to act effectively after the fashion of a beam for loads of great span. The orogenic structures, on the other hand, give maximum stresses much nearer the surface, in the stronger lithosphere; because of their shorter wave-lengths they do not produce in it bending stress as in a loaded beam and affect comparatively little the deeper-seated asthenosphere.

If, then, it is known from the preceding theoretical considerations that the limits of strength of the lithosphere and asthenosphere determine the limits of the departures from isostasy, the analysis of the nature of these departures may lead in turn to a knowledge of the distribution of strength.

CHARACTER OF THE CURVE OF STRENGTH

In curve F′ of Fig. 18 is shown the nature of the curve of strength as suggested by the geodetic evidence from the United States. In curve F″ is shown the nature of the curve as suggested by the departures from isostasy exhibited by the great mountain axes and possibly by the continental ice sheets. These curves may be taken as showing the value of the elastic limit at various depths for permanent stresses. With varying geologic conditions, especially those connected with rising magmas and their emanations, the curve of strength must vary widely, and furthermore no very close parallelism of strength-curve and stress-curve is to be expected. These curves, therefore, are intended to bring out general relations; they are of qualitative, not quantitative value. The drawing of curve F″ somewhat inside of curve E means that below the point of maximum stress in E, as given for a homogeneous elastic earth, the stress is assumed as somewhat greater than the crust at those levels can sustain. Upon the development of this load plastic flow in these deeper levels would take place maintaining the stress within the strength curve for each level; the crust above would come to act to some extent as a bending plate, the stresses within it would increase, chiefly within the upper and lower portions. This added strain would compensate for the yielding below. For the reasons discussed previously, however, showing the structural weakness of the lithosphere as a beam, this action, it is thought, could not go very far, and, in consequence, the loads on the lithosphere are essentially such as to give stresses contained within it, distributed according to Darwin's law. The preceding deals only with that part of the curve of strength which marks the gradation from lithosphere to centrosphere. The relations of this part to those above and below need still to be considered.

The highest stress found for the loads regarded as harmonic waves was for the great folds on the floor of the Pacific Ocean. These were taken as equivalent to harmonic waves of rock of density 2.67, 400 km. in wave-length, and 4,000 m. in amplitude. But even these folds give a maximum stress of only 393 kg. per sq. cm. (5,590 pounds per square inch), and this at a depth of 64 km. At the surface strong limestone or granite can

sustain a stress-difference of 1,750 kg. per sq. cm. (25,000 pounds per square inch), and selected specimens show ultimate breaking strength approaching 2,800 kg. per sq. cm. For stresses of geologic endurance and in the heterogeneous outer crust it is probable, however, that stress limits should be chosen below 1,750 kg. per sq. cm.

The work of Adams and King has shown that small cavities in granite are not closed when the rock is subjected to the pressure and temperature normally existing in the earth at a depth of 11 miles.[1] The presence of occluded gases acting through great lengths of time, by facilitating recrystallization, might affect this result of laboratory experiments, but the capacity of dry rock to sustain even greater cubic pressures without yielding seems to make safe the conclusion that except in the presence of magmatic emanations the crust at a depth of 11 miles (17.7 km.) is able to bear a stress difference of 100,000 pounds per square inch and is at least four times as strong as rock close to the surface.

At twice this depth, however, the temperatures become such that if it were not for the great pressures even dry rocks would approach a molten condition. The presence of high temperatures and of gases which may act as crystallizers presumably becomes dominant at such depths over the effects of the increasing pressures. We many conclude, therefore, that the maximum strength of the crust in regions free from igneous activity is found at levels above rather than below 40 km. and may lie between 20 and 30 km. deep.

To bring to a focus this discussion a tabulation of ratios of strengths for increasing depths may be given, as derived from the strength curves F′, F″ of Fig. 18, the standard being taken as the strength of surface rocks. By giving them merely as ratios and stating that the average strength of the solid rocks at the surface is itself an uncertain quantity owing to complications of structure and composition, the appearance of an undue certainty is avoided.

The general conclusion which stands out from this tabulation is that the weakest part of the asthenosphere is of the order of one one-hundredth of the maximum strength of the lithosphere and is perhaps only a twenty-fifth of that of massive surface rocks. Its

[1] *Jour. Geol.*, XX (1912), 97–138.

limit of capacity for sustaining stress-differences is apparently of the order of 1,000 pounds per square inch, though its weakness may be masked to some extent by the strength above. From the evidence, however, it seems capable of carrying stresses of more than 100 pounds per square inch, but is clearly incapable of carrying stresses of as much as 5,000 pounds per square inch. To reach a

TABLE XXX

ESTIMATED APPROXIMATE RATIOS GIVING THE VARIATION OF STRENGTH WITH DEPTH AS SHOWN BY THE NATURE OF DEPARTURES FROM ISOSTASY

LITHOSPHERE

Depth in Kilometers	Strength in Percentage
0	100
20	400
25	500
30	400
50	25
100	17

ASTHENOSPHERE

Depth in Kilometers	Strength in Percentage
200	8
300	5
400	4

more definite conclusion the subject must be tested from many angles and is a problem for the geophysicist rather than for the geologist, but the results are of geological importance and the geologic and geodetic data may turn out to have more determinative value on the distribution of strength than the evidence from tides and earthquakes.

[*To be continued*]

SOME ELLIPSOIDAL LAVAS ON PRINCE WILLIAM SOUND, ALASKA[1]

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Many descriptions of ellipsoidal, spheroidal, or pillow lavas have been published from time to time. Rocks of this type have a wide geographic range and have been formed at intervals from pre-Cambrian to the present. Various conclusions have been reached by different writers as to the mode of origin of ellipsoidal lavas, and the reasons for the development of their peculiar ellipsoidal, spheroidal, or pillow-shaped forms. The ellipsoids or pillows have been considered to be (1) concretionary, (2) the result of brecciation, (3) due to contraction, (4) accumulations of volcanic bombs, (5) the result of explosive eruptions, (6) and the result of submarine cooling. In a recent article N. Sundius[2] has well summarized the important literature on the subject.

Notwithstanding the various hypotheses which have been advanced to explain this phenomenon, it has come to be generally accepted that the ellipsoids are the result of the flowing of the lavas into water, or their extrusión under water, and are due to the rapid cooling of the flowing molten material under these conditions. Tempest Anderson[3] has described recent pillow lavas actually seen in the process of formation, and there now seems to be little reason to doubt that the chilling of lavas under water is responsible for many of the ellipsoidal lavas with which we have become familiar. The purpose of this paper is to present certain facts which seem to the writer to prove conclusively that the ellipsoidal lavas near Ellamar, Alaska, were subaqueous flows, and to give certain criteria

[1] Published by permission of the Director, United States Geological Survey.

[2] N. Sundius, "Pillow-Lava from the Kiruna District," *Geol. fören. Förhandl.*, Bd. 34, Häft 3 (1912), pp. 317–33.

[3] Tempest Anderson, "The Volcano of Matavanu in Savii," *Quarterly Jour. Geol. Soc.*, XLVI, No. 264 (1910), 621–39.

which, in this field at least, proved of value in determining which was the upper and which the lower side of a given flow as deposited, thus aiding in deciphering the structure in a region where the normal attitude of the beds had in many places been obscured, and where overturning of the beds might be expected.

The ellipsoidal greenstones here described were studied in the vicinity of Ellamar, Alaska, in the fall of 1912, during an investiga-

FIG. 1.—Cross-section of lava flow, showing ellipsoidal forms. The ellipsoids are characterized by both radial and concentric sets of cracks. Photo. by S. Paige.

tion of the geology and mineral resources of the district by Mr. B. L. Johnson and the writer. The exact age of the greenstones is not known, but they are associated with a thick series of sedimentary beds which are believed to be for the most part of Mesozoic age. The greenstones were of especial interest because all of the copper deposits of the district are in the greenstone, or closely associated with it. In those areas in which the greenstones are most highly mineralized the lavas have been extensively sheared and meta-

morphosed, so that they have almost completely lost their original appearance. In other places, however, the lavas have been tilted, with only slight folding and with little shearing, so that their original character is well preserved. The term greenstones has been commonly applied to these rocks, which are all diabases or basalts. The writer prefers the term "ellipsoidal" to other commonly used terms such as "spheroidal" or "pillow" lavas, for the forms are

FIG. 2.—Uptilted ellipsoidal flows, showing irregular shapes of the flow forms. Photo. by B. L. Johnson.

most often ellipsoidal in cross-section (Fig. 1), but in plan they appear as more or less continuous bodies with irregular bulges and protuberances (Figs. 2 and 3).

The best exposures for studying the ellipsoidal lavas were found along the shore from two to four miles northwest of the village of Ellamar, toward Rocky Point. There the waves have cut steep sea cliffs which rise from the water's edge, often with no beach at all, and the exposures are fresh and perfectly free from covering.

An examination of this section shows that although the ellipsoidal flows form the larger proportion of the series, they are interbedded with some more massive diabase beds. At intervals, however, the extrusive rocks are seen to be interrupted by slates and graywackes, the sedimentary beds becoming thicker and more frequent as the top of the greenstone series is approached. The strike and dip of adjoining lavas and sediments is the same, and it

FIG. 3.—Ellipsoidal lavas. The forms are ellipsoidal in cross-section, but in plan are long uneven bodies characterized by bulbous portions separated by constrictions. Photo. by S. Paige.

is evident that clastic sediments were deposited at intervals between the outpouring of the lava flows (Fig. 4). In a large number of observations on slate or slate-graywacke beds interbedded with ellipsoidal greenstones, no single case was found in which there was a discordance in bedding between the sediments and the lavas. Furthermore, in the cracks between the ellipsoids of the lower part of a flow and in the smaller cooling cracks within the individual ellipsoids there were frequently seen thin vein-like layers of a black, shaly material which proved to be of the same character as the under-

lying sediments. The presence of this sedimentary material in the cracks of the fractured ellipsoids can be satisfactorily explained only by assuming that the bed of mud upon which the lava was extruded was at that time soft and plastic, and that upon the cooling of the lavas, and their cracking from shrinkage as they chilled, the weight of the lava bed forced the soft mud from below up into the cracks of the lava as fast as they were opened. In places these

FIG. 4.—Steeply dipping ellipsoidal lava, on right, succeeded by conglomerate and graywacke. The uneven lower surface of the conglomerate conforms to the top of the underlying lava flow.

mud-filled cracks extend several feet into the flow, and where best developed give the rock a mosaic appearance, the fragments of greenstone, often only a few inches in diameter, being surrounded by a black layer of baked shale. The mud fillings of the cracks vary in width from the thickness of a knife blade to several inches. At one locality the openings between the ellipsoids were found to be filled with limestone.

In those places where thin ellipsoidal greenstone beds were adjoined both above and below by sedimentary beds, it was found that the two surfaces of the flow presented different appearances (Figs. 4 and 5). The under surface in each case was flat, the ellipsoids of the lower layer being flattened at the base, though presenting

curved surfaces on their other sides. At one place a steep overhanging sea cliff 30 feet high was formed by the base of such a flow, the underlying slate having been cut away. The base of this flow was flat and smooth, though the flow, as seen in section, was composed altogether of ellipsoidal lavas. Into the cracks in the base of this flow the mud had penetrated deeply. It is believed that the flat lower surface of this ellipsoidal flow could have been

FIG. 5.—Ellipsoidal lavas below, separated from more massive diabase above by an intercalated shale bed. The lower surface of the shale is uneven, having taken the shape of the ellipsoidal lava surface upon which it was deposited. The overlying flow is flat on the bottom, conforming to the shale surface. Photo. by S. Paige.

formed only by lava which was poured out upon a flat bed of sediments, the lava having conformed to the surface upon which it was deposited. The upper surface of an ellipsoidal bed presents a different appearance (Figs. 4 and 5). There the flow surface was under the pressure only of the column of water above it, and the surface hardened to form an irregular floor consisting of a succession of domes, like the surface of a magnified cobblestone pavement, the surface of the ellipsoids representing the cobble-

stones. Upon this surface later deposits of mud were laid down, the bottom of the mud bed taking the shape of the irregular lava surface. The lava flows show no signs of weathering or erosion upon their surface, and the succeeding mud beds must have been laid down soon after the lavas cooled.

Summary.—The greenstones near Ellamar, Alaska, comprise a thick series of lava flows, most of the flows consisting of ellipsoidal lavas, and many of them displaying this character in a striking way. At frequently recurring intervals in the greenstone series there are interbeded slates and graywackes, and rarely some limestone, the sedimentary beds being in every case structurally conformable with the overlying and underlying lavas. The surfaces of the flows show no evidence of erosion or weathering, the succeeding sediments having been laid down upon the fresh, unaltered lava surface. If the greenstones are subaqueous flows, as is believed to be the case, then they were merely interruptions during the deposition of ordinary clastic sediments.

If, on the other hand, the lavas were not subaqueous flows, and were extruded upon land, it is necessary to postulate an elevation of the surface above sea level for each extrusion of lava and an immediate resubmergence for the deposition of the sedimentary beds. In view of the many recurring sedimentary beds throughout the greenstone series, this latter assumption seems most unlikely. The absence of evidence of erosion or weathering on the surface of the lava flows also supports the conclusion that they solidified below sea level, as does the presence in the lava cracks of thin films of mud which must have been plastic at the time they were injected.

The difference in appearance of the flat bottom of an ellipsoidal flow, with its abundance of mud-filled cracks, from the uneven upper surface of the same flow, and the consequent unevenness of the bottom of the succeeding sedimentary bed often made it possible to determine which surface of a steeply tilted bed was originally the upper surface, and was a valuable aid in working out the structure of the beds.

A SKETCH OF THE LATE TERTIARY HISTORY OF THE UPPER MISSOURI RIVER

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The story of the Missouri River forms one of the most interesting chapters in the geologic history of the Great Plains. The valley of the present stream is composed of many parts, some relatively young and others moderately old, but as a unit it is youthful when compared with the valley of the Mississippi. Notwithstanding its youth, its history is so complex and so little known that only a mere sketch can be given at this time. All that will be attempted, therefore, will be to point out a few late Tertiary river channels, which may indicate former courses of the Missouri River and some of its larger tributaries, and sketch their bearing on the history of the main stream. One of these stream courses has been traced from Poplar, Montana, northeastward to the North Dakota boundary. Another valley, partially abandoned, extends northward from a point 6 miles west of the mouth of Bowlin Creek on the Little Missouri River to the mouth of Tobacco Garden Creek on Missouri River (see map, Fig. 1).

F. H. H. Calhoun[1] has mapped a number of preglacial valleys in north-central Montana, many of which are connected with the history of the Upper Missouri. One of the most important of these valleys extends from the mouth of Little Sandy Creek northeastward for about 35 miles to Milk River near Havre. Below its junction with this old valley the valley of Milk River is 2–2½ miles broader than above and clearly indicates that the headwaters of a larger stream, probably the Missouri, emptied into it and followed its course previous to the ice invasion. East of the mouth of Little Sandy Creek the present Missouri flows in a narrow, rocky channel for many miles and joins the old channel at the

[1] F. H. H. Calhoun, "The Montana Lobe of the Keewatin Ice Sheet," *U.S. Geol. Survey, Prof. Paper 50.*

mouth of Milk River, about 20 miles below Glasgow. Another old valley, noted by the writer in 1911, extends from Poplar, Montana, northeastward to Medicine Lake, thence eastward for several miles, thence northward to the vicinity of Dagmar and Coalridge, and thence northeastward to the lake region of the extreme northwestern corner of North Dakota. Beyond this the old river channel is completely buried by drift.

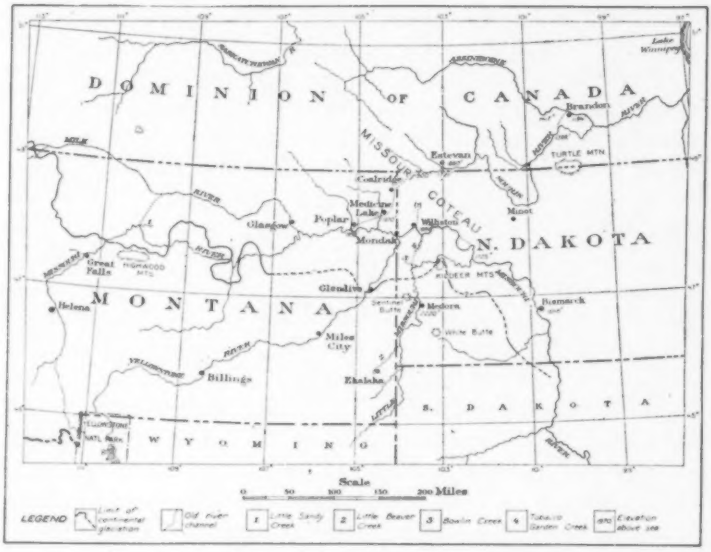

FIG. 1.—Map of Upper Missouri River, showing some Tertiary river channels. By C. M. Bauer.

In the vicinity of Poplar this old valley is nearly filled with glacial drift. The town of Poplar is built on a moraine, as shown by the section of drift 45 feet thick just north of the Indian School on Poplar River. Northeast of Poplar for several miles the drift is probably much thicker. In T. 28 N., R. 52 E., and northeastward for 20 miles the trough of the old valley is from 3 to 4 miles wide, with broad, gentle slopes on either side. Near Medicine Lake it widens, and on its south bank there is an area of sand dunes.

Small lakes are common along the old channel, and in the vicinity of Coalridge the valley trough is very distinct. A. L. Beekley traced this old valley from Medicine Lake to the North Dakota boundary in 1910, and describes its features in *Bulletin 471*, U.S. Geological Survey.

From the Missouri Coteau northeastward the ancient stream course is unknown, but it seems probable that it joined the preglacial Souris River and flowed thence into the predecessor of Assiniboine River, thence northward by way of Lake Winnipeg and Nelson River to Hudson Bay.

Elevations above sea along this course, some of which are shown on the map, point to this conclusion. Most of the altitudes given are of the lowest point in the immediate locality. If we take these altitudes in order, beginning at Poplar, Montana, and continuing northeastward, we find that, with the exception of Medicine Lake, which is 1,970, there is a gradual decrease in altitude from 1,958 feet on the present Missouri River bottom to 1,194 feet at Brandon on the Assiniboine River. It is significant that the present fall of the Missouri River in an equal distance is much less. The fact that Medicine Lake is now 12 feet higher than the flood plain at Poplar is probably due to postglacial erosion by the Missouri River. Another factor which must be considered is that the altitudes given do not indicate definitely the position of bedrock because of the varying thickness of glacial drift and alluvium in the old channel. It is also possible that differential warping to the amount of 200 feet, known to have occurred between latitudes 47° and 51° N., in the area occupied by Lake Agassiz and Lake Souris affected this region also and has relatively raised the country to the north from what it was in late Tertiary time.[1]

Changes in Yellowstone River due to glaciation are apparently few. Its present course, to within a few miles of its mouth, is about the same as it was in preglacial time. However, its anastomosing channel from Glendive to Mondak, on a flat, sandy, valley floor, is evidence of glacial filling in this portion of its channel. From a few miles south of Mondak the Tertiary Yellowstone probably fol-

[1] Warren Upham, "The Glacial Lake Agassiz," *U.S. Geol. Survey, Mon. 25*, pp. 474–522, 1896.

lowed the present course of the Missouri River to Williston, thence northward along the valley of Muddy Creek, which is much too wide for the stream that now occupies it. The writer has not traced this portion and was unable to get definite information on the old valley of the Yellowstone more than a few miles north of Williston.[1] However, Lambert, in *Exploration for the U.S. Pacific Railroad in 1853*, mentions a wide valley, discovered by Lander, which crosses the Coteau connecting the Souris with the Missouri near old Fort Union (Williston?) It seems probable, therefore, that the ancient stream proceeded north or northeastward from this point.

Concerning the course of the Tertiary ancestor of the Little Missouri River, the evidence is very plain. An old valley, several miles in width, extends northward from the first prominent eastward bend of the present stream near the mouth of Bowlin Creek to the head of Tobacco Garden Creek and thence along this valley to the present Missouri and probably joins the Tertiary Yellowstone near Williston. This old valley south of the Missouri River was first noted by F. A. Wilder,[2] who describes it as being evidently a former course of a prominent stream similar to the Little Missouri.

The present valley of the Little Missouri River from the Kildeer Mountains east to its mouth is narrow and bordered by rugged badlands of soft Fort Union strata. Consolidated glacial drift occurs on the jagged tops of some of the ridges 250 feet high and within one-half mile of the stream channel, but nowhere on the sides of the valley has consolidated drift been found. Although there are no terraces in this portion of the valley, in the vicinity of Medora and continuing southward into South Dakota a well-defined terrace from a few feet to several miles in width has been mapped by C. J. Hares.[3] Along Little Beaver Creek, one of the tributaries of the Little Missouri, the remaining fragments of this

[1] F. H. H. Calhoun, "The Montana Lobe of the Keewatin Ice Sheet," *U.S. Geol. Survey, Prof. Paper 50*, pp. 35, 36, 1906.

[2] F. A. Wilder, "The Lignite of North Dakota and Its Relation to Irrigation," *U.S. Geol. Survey, Water Supply, Paper 117*, p. 43, 1905.

[3] C. J. Hares, "Lignite in Southwest North Dakota," *U.S. Geol. Survey Bull.* (in preparation).

terrace have been mapped by the writer nearly to its headwaters. A few miles west of Ekalaka, Montana, several flat-topped, gravel-covered ridges 50 to 60 feet high form a prominent feature of the landscape. The terrace, where noted, is covered with gravel from a few inches to 10 or 12 feet in thickness. The gravel consists principally of red and white quartzite pebbles as well as some of chert and argillite, ranging from the size of sand grains to 2 or 3 inches in diameter. The terrace has its best development along the Little Missouri in southwestern North Dakota, and at this place it probably represents the late Tertiary valley floor of the Little Missouri, which existed until the invasion of the earlier ice sheet. Since then the Little Missouri has cut a channel to the east and shortened its course. The present trough of the Little Missouri, which is about 200 feet deep near Medora, is therefore the result of erosion since the disappearance of the early ice sheet.

Parts of the Missouri River above Poplar, Montana, and likewise the Yellowstone River, probably had their beginning at the end of the Cretaceous period, when the seas withdrew and exposed large areas in the region of the Rocky Mountains to stream erosion. These primitive streams and their associates were apparently small at first, but as the land area grew by withdrawal of the sea and with the rising of the mountain belt, in the early part of the Eocene epoch they became larger, gained velocity, and gathered sediment which was gradually spread out over the region now known as the great Fort Union area. These rivers doubtless meandered widely over this broad palustral flat and had their *embouchure* somewhere to the north,[1] probably into the predecessor of Hudson Bay or the Arctic Ocean.

As the primitive Yellowstone and Upper Missouri and their tributaries and allies continued their work, several thousand feet of sediment were deposited during the Eocene epoch in eastern Montana, western North Dakota, and parts of adjoining states. This was finally interrupted by an epeirogenic movement, which

[1] W. G. Tight, Abstract of paper read before the Geological Society of America, *Geol. Soc. Amer. Bull.*, XVIII.

Warren Upham, "Age of Missouri River," *American Geologist*, XXXIV (1904) 80–87.

caused a relative rise of this portion of the continent, together with some folding. The streams were now set for the task of erosion. During this time the Yellowstone succeeded in maintaining its course across the Glendive anticline, a long, narrow arch extending southeastward from the Yellowstone River, Montana, to the northwestern corner of South Dakota, and the Upper Missouri maintained its course across the Poplar dome, a low quaquaversal structure at Poplar, Montana.

Deposition again began on the Great Plains in Oligocene time.[1] The change in level causing deposition may have been the result of warping, the mountains on the west rising relative to the plains and furnishing more sediment.

A general uplift following the Oligocene inaugurated a great period of erosion and base-leveling on the northern Great Plains. In places, particularly in Saskatchewan,[2] the great rivers deposited some gravel and coarse sediment where the gradient became low and the load too great. However, during the late Tertiary, the Upper Missouri and Yellowstone carried away thousands of feet of strata and carved for themselves magnificent broad valleys in the soft Cretaceous and early Tertiary deposits.[3] In the vicinity of Highwood Mountain the erosion in the valleys probably amounted to about 3,000 feet. Farther east, near Sentinel Butte and the Kildeer Mountains, about 1,000 feet and near Turtle Mountain about 500 feet of material was excavated. Between these monadnocks the country was reduced to a peneplain. Then, according to Upham, followed the early Quaternary uplift and the glaciation of the northern part of the continent. The outlets to the north became blocked with ice and later were partly filled with drift and outwash. Ponding of waters and marginal drainage began. Lake Agassiz and Lake Souris were formed, as well as many other smaller lakes, most of which are now extinct. Just where the

[1] A. G. Leonard, geologic map of North Dakota, 1913, showing present distribution of White River formation.

[2] R. G. McConnell, "On the Cyprus Hills, Wood Mountain and Adjacent Territory," *Canada Geol. Survey Report*, N.S., I (1885) 70–c.

[3] Warren Upham, "Tertiary and Early Quaternary Base-leveling in Minnesota, Manitoba, etc.," *American Geologist*, XIV (1894), 235–46.

marginal drainage proceeded when the ice was at its maximum in eastern Montana and western North Dakota is at present a matter of conjecture. However, it is probable that it flowed southeastward across cols and through various preglacial valleys until it joined the Mississippi.[1]

After the retreat of the ice, the Upper Missouri, the Yellowstone, and the Little Missouri, owing to the glacial débris, no longer found an outlet to the north, but instead followed in part the glacial drainage and in part the preglacial stream courses southeastward to the Gulf of Mexico.

[1] J. E. Todd, "The Pleistocene History of Missouri River," *Am. Assoc. Adv. Sci. Proc.*, N.S., XXXIX (1914), 263–74.

THE INTERGLACIAL GORGES OF SIX MILE CREEK AT ITHACA, NEW YORK

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University of Illinois

INTRODUCTION

The question of the complexity of the glacial period in central New York is still largely unsettled. It is conceded probably by all that the region has suffered more than a single ice invasion, but how many such invasions there were and how far the different invasions were separated from each other in time are problems as yet only partially solved.

The criteria for the identification of distinct glacial epochs which can be applied in a hilly region well within the southern limits of the ice, and even within the zone of active erosion by the latest Wisconsin ice sheet, are necessarily different from those which have been used so extensively and so successfully in the Mississippi Valley. Under the conditions prevailing in central New York one could not expect, for instance, to find drift sheets imbricated in the same manner as on the plains of the Mississippi near the limit of glaciation where the erosive power of the ice was doubtless slight and undirected and where deposition was the rule. Nor would it be possible to apply the common test of the relative erosion suffered by drift sheets of different ages, for, since the region in question lies within the great moraines of the Late Wisconsin epoch, any erosional features developed on earlier drift sheets would have been modified or destroyed by the Wisconsin ice. The hilly nature of the country, too, would tend to prevent the accumulation of till in even, continuous sheets and would lead to an irregular, patchy distribution of any remnants of earlier drift deposits. The meagerness, in this region, of the evidences from glacial deposits indicating multiple glaciation may be judged from the fact that, so far as the writer has been able to discover, only one investigator, Carney,[1] has

[1] "Pre-Wisconsin Drift in the Finger Lake Region of New York," *Jour. Geol.*, XV (1907), 571–85.

definitely claimed to have found such deposits. Tarr[1] mentions two or three exposures of possible older drift near Ithaca, but believes that the evidence is inconclusive, and Gilbert[2] recognizes the possibility of two till sheets and the certainty of "an ep :h of local till erosion by a glacier," but is led to no definite conclusions further than this.

There is, however, another criterion of multiple glaciation for the application of which the Finger Lake valleys furnish ideal conditions. This criterion is the amount of work performed during interglacial intervals by streams whose normal profiles have been thrown out of harmony by glacial erosion or deposition; streams which must, therefore, resort to gorge-cutting in their endeavor to bring their beds back to normal grade.

A glacier moving along a valley tends to overdeepen it and to leave tributary valleys hanging. As soon as the ice withdraws, the streams in the hanging tributary valleys begin the work of bringing their beds down to the normal profile of equilibrium. In doing this they cut gorges in the lower ends of the hanging valleys. Given sufficient time after an ice invasion, the tributary streams would enter the main valley at grade through narrow, gorgelike valleys cut in the bottoms of the hanging valleys. These gorges, narrow at first, would, after their streams have reached grade, continue to widen in a normal manner under the influence of the ordinary weathering agencies combined with lateral swinging of the stream.

After a single glacial epoch the streams in all the hanging valleys would be flowing in such gorges and would, in time, bring their beds down to a grade accordant with that of the main stream. Suppose, now, a second glacial epoch should intervene. The chances are that the main valley would be deepened still further by ice erosion while the gorges in the tributary valleys would become more or less clogged with drift. On the withdrawal of this second ice sheet the gorge bottoms of the tributary valleys would be left hanging above the bottom of the recently deepened main valley.

[1] Watkins Glen-Catatonk Folio, No. 169, U.S. Geol. Survey.

[2] G. K. Gilbert, "Boulder Pavement at Wilson, N.Y.," *Jour. Geol.*, VI (1898), 771–75.

Should a tributary gorge have been incompletely filled with drift, its stream would begin the task of clearing out the drift and of cutting a second gorge in the bottom of the earlier one. Should it happen, however, that the drift filling of a gorge was so complete that the stream on reoccupying its old valley and taking a consequent course found lower ground to one side, a new and independent gorge would be cut. Should the consequent course induced by the drift topography correspond in part with that of the old gorge and differ from it in part, the stream would cut a new gorge where it found itself out of the old channel, and merely clear out the old gorge where the two courses were in coincidence.

Parts of the old drift-filled gorge unoccupied by the stream would be preserved and would remain indefinitely as fossil gorges hanging above the level of the main valley bottom.

If an interval of deglaciation sufficiently long ensued, the tributary streams would again bring their beds down to grade. A third epoch of glaciation would bring about a repetition of events with the possible formation of still another gorge in the hanging valley bottom. The number of distinct gorges should, then, give at least the minimum number of glacial invasions which the region had suffered. It would not necessarily give the maximum, for during one or more of the intervals the stream might have merely re-excavated one of the older gorges.

The relative width of gorges cut during different intervals should, other things being equal, give a rough measure of the relative length of the interglacial intervals which they represent. A comparison of the width of the older gorges with that of the postglacial gorge should, in the same way, give a rough measure of the length of the interglacial interval as compared with postglacial time. This would be true, of course, only providing the older gorges were occupied by streams during only one interval.

The criterion outlined above for the determination of repeated ice invasions is considered to have advantages over others in that the possibility of minor halts and readvances of the glacier being interpreted as distinct glacial epochs is reduced to a minimum, for streams require time for the excavation of gorges. If an interglacial gorge is as large or larger than the postglacial gorge of the same

stream, it is a fair inference that the interglacial interval bore a similar ratio to postglacial time. It would be necessary, however, to take account of possible different gradients of the streams and different climatic conditions during the intervals, but the effects of these could hardly lead to the confusion of true interglacial intervals with minor retreats of a fluctuating ice front.

The conditions outlined in the preceding paragraphs are realized very fully in the Finger Lake valleys. The main troughs, extending approximately in the direction of ice movement, were deeply scoured, while tributary valleys, suffering less erosion, were left hanging. The extent to which hanging valleys were developed has been so well presented in the writings of Tarr and others that it need not be dwelt upon here.

If the glacial period in this region was represented by more than one ice invasion, with corresponding intervals of deglaciation, we should expect to find, in at least some of the numerous tributary hanging valleys, old drift-filled gorges representing the work of interglacial streams. Such gorges are abundant, though their interglacial origin has not, in most cases, been fully demonstrated. They have been described by Matson[1] as occurring in the Buttermilk valley, and by Tarr, in the valleys of Watkins Glen, Fall Creek, Six Mile Creek, and others in the neighborhood (see references in later paragraphs).

In the autumn of 1909 the senior author discovered in the valley of Six Mile Creek phenomena which apparently had not been recognized before and which seem to have a distinct and important bearing on the problem of multiple glaciation. During the year 1910–11 Mr. Edwin A. Filmer, at the senior author's suggestion, undertook a detailed study of the lower Six Mile Creek valley. The results of this study were embodied in a paper now in manuscript form in the Cornell University Library. The facts brought to light by our investigation of the valley seem worthy of wider publicity than is possible for this manuscript. Hence the present joint paper has been prepared by the senior author, who has, however, made free use of Mr. Filmer's material and conclusions.

[1] G. C. Matson, "A Contribution to the Study of the Interglacial Gorge Problem," *Jour. Geol.*, XII (1904), 133–51.

THE VALLEY OF SIX MILE CREEK AND ITS GORGES

The troughlike Cayuga Lake basin splits at Ithaca into two prominent prongs, the one carrying the Inlet stream from the southwest and the other occupied by Six Mile Creek, which enters from the southeast (see Fig. 1). Of the two, the Inlet valley is the deeper

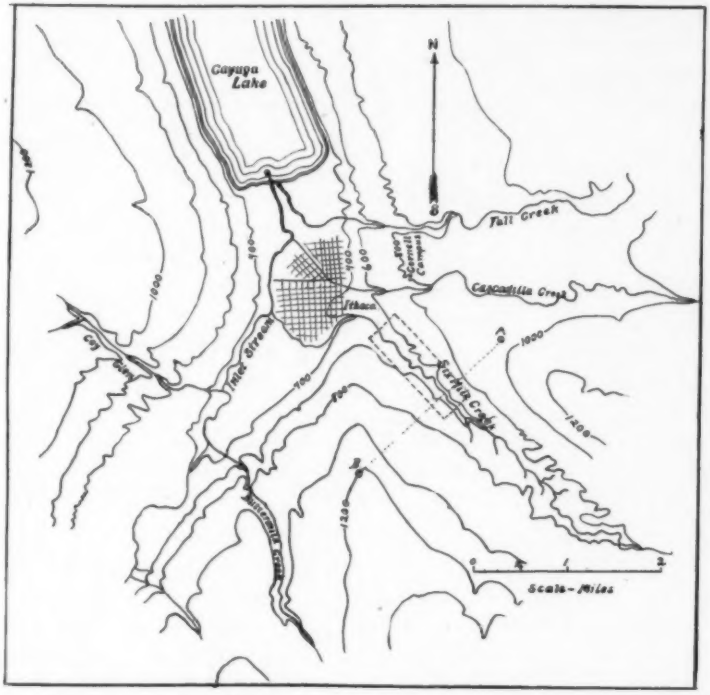

FIG. 1.—Map of the Ithaca region, showing the course of lower Six Mile Creek and its relation to the Cayuga valley and to other streams in the neighborhood. Area shown on Fig. 2 is inclosed by dotted lines.

and the steeper walled and is the main prolongation of the lake basin. It continues with a gradual rise of the valley bottom for about 10 miles southwest of Ithaca and then, by a "through valley" connects southward with one of the tributaries of the Susquehanna. It is a typical glaciated trough. Tributary valleys along this

trough are hanging and their streams enter through narrow rock gorges.

The Six Mile Creek prong is similar to the Inlet trough though it is not so deep and its sides are more flaring. The valley extends southeast from Ithaca in almost a straight line to Caroline Depot, about 7 miles to the southeast, where a southern prong of the valley connects by a low "through valley" past White Church and Willseyville with one of the headwater branches of the Susquehanna leading directly south.

That the valley of Six Mile Creek and this "through valley" were channels of vigorous ice motion during the glacial period is attested by (1) the character of the "through valley" at Willseyville, (2) the trunkated spurs and straight walls of the valley round White Church and Willseyville, (3) the distribution of the moraines in this latter valley and that of Six Mile Creek,[1] and finally by (4) the boat-shaped, typically glaciated cross-section of the lower Six Mile valley, recognized first by Simonds in 1877[2] and later by Tarr.[3] Simonds says (p. 51), referring to the Inlet and Six Mile Creek valleys:

> These deep, well-worn valleys are undoubtedly the result of glacial action. The mass of ice which filled the Cayuga Lake basin divided at its southern extremity. One part, the larger, flowed to the south, wearing down the Inlet valley, and the other traversed the Six Mile Creek valley, both of which were occupied by preglacial streams.

And Tarr (*op. cit.*, p. 20, first paragraph) says:

> . . . . Both Salmon and Six Mile Creek valleys hang at a much lower level than their neighbors (for example Fall, Cascadilla, and Buttermilk [Ten Mile]). I am now convinced that the interpretation of this discordance as opposed to the glacial erosion theory was incorrect and that these two valleys are really confirmatory of the glacial erosion theory. This change in view is the result of a recent study of the valley profiles and a mapping of the morainic deposits of the valleys in question. The latter show that these valleys were occupied by actively moving ice parallel to their axes while the neighboring higher hanging valleys were not. A study of the profiles shows that these discordant hanging valleys have the U-shape of glacial erosion and not the gorge

[1] See Tarr, Watkins Glen-Catatonk Folio.

[2] Simonds, *American Naturalist*, XI (1877), 49–51.

[3] *Jour. Geol.*, XIV, No. 1 (1906), 20.

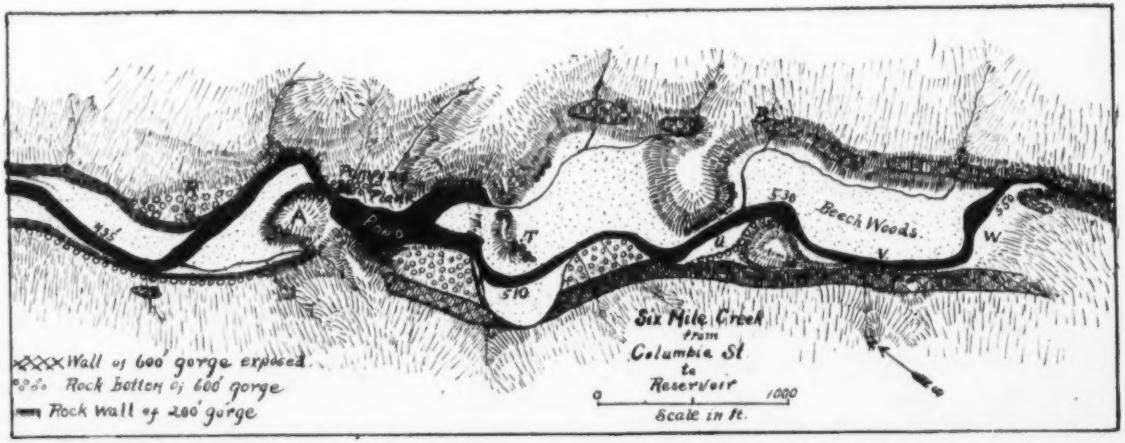

FIG. 2.—Map of a portion of the course of lower Six Mile Creek, showing the topography, the buried gorges, the outcrops of the rock walls and bottoms of the earlier gorges, and the elevations of the stream bed in different parts. The stream flows from right to left of the map. This map is reduced and somewhat modified from a map of the Six Mile Creek valley prepared by the city of Ithaca in connection with surveys for a water supply. Elevations shown represent height above sea-level and were taken from this map. They are approximately correct, but the writers cannot vouch for absolute accuracy.

shape of a rejuvenated valley, the only other explanation which seems a possible one for such discordance. It is believed, therefore, that while the Cayuga valley was profoundly deepened by ice erosion the Salmon and Six Mile Creek valleys were deepened moderately and the Fall, Cascadilla, Buttermilk, and other valleys practically not at all, since they were not occupied by ice freely moving along their axes.

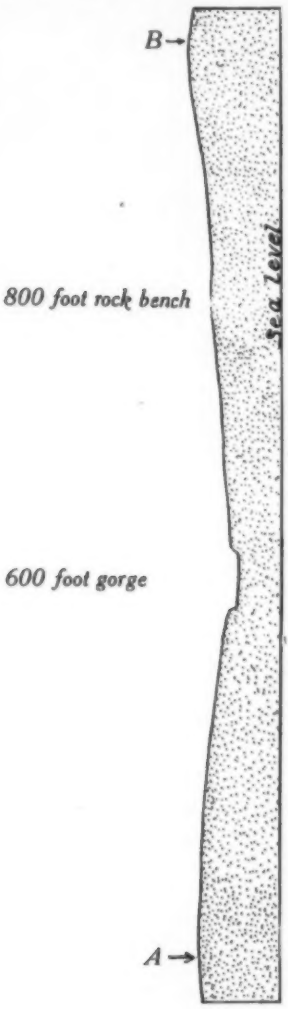

FIG. 3.—A profile across Six Mile Creek valley from *A* to *B* of Fig. 1, drawn to natural scale—vertical and horizontal scale the same (2,000 feet to one inch). This profile shows the relation of the 600-foot gorge to the broad valley of Six Mile Creek.

The evidence cited above, together with the foregoing quotations, makes clear the point which we especially desire to emphasize, namely, that the Six Mile Creek valley was occupied by actively moving ice and that it is hanging at a lower level than others near by because glacial erosion deepened its bottom, forming the U-trough so well shown in the profile (Fig. 3).

The rock bottom of the U-trough of lower Six Mile Creek valley is hanging at an elevation about 100 feet above the level of Cayuga Lake, and about 450 feet above the rock bottom of the Inlet valley at this point.

In the bottom of Six Mile Creek trough there is a series of three distinct gorges, which, according to our interpretation, are of different ages. Two are partly, and in places entirely, filled with glacial drift; the third is postglacial. The characteristics of these gorges and their relations to each other

will appear from the following description taken in connection with the map (Fig. 2). There is an older gorge (hereafter referred to from its general width as the 600-foot gorge), 500–600 feet wide, of low gradient, in the bottom of which a second much narrower gorge, which we will call the 200-foot gorge, has been cut to a base-level some 100 feet lower than the other. Both of these gorges antedate the last ice invasion, being still filled by glacial deposits except where they have been cleared out by the postglacial stream.

A third, much younger, gorge (postglacial) is found in places where drift and lake delta deposits have so completely filled the earlier gorges that in re-excavating its valley the stream took a new

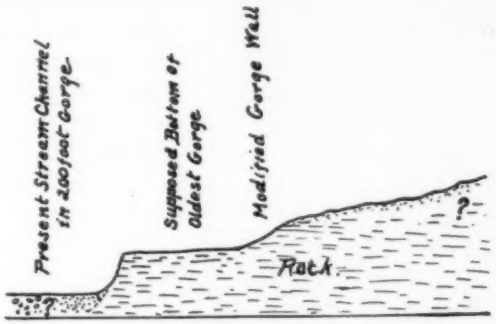

FIG. 4.—Sketch showing bottom and sides of gorge as it appears from upstream

course which, in this case, did not correspond with the former one. In cutting down in this new course the stream, encountering rock, developed the postglacial gorge.

The bottom and walls of the oldest, or 600-foot, gorge are exposed in numerous places for a distance of over a mile. The first evidence is seen just below the Ithaca waterworks pumping plant, where a rock bench, representing the old gorge bottom and east wall, shows in section where the postglacial stream has been undercutting its banks. The characteristics of this section as seen from upstream are indicated in the accompanying sketch (Fig. 4). A little farther upstream, just below the road, is an interesting section best illustrated on the map (Fig. 2). A rock island (*A* of Fig. 2) is found between the buried 200-foot gorge on the south and the recent postglacial gorge. The top of the rock in this island lies approximately

at the level of the previously described rock bench across the stream just below. The rock bench and the rock island together are interpreted as the bottom of the oldest, or 600-foot, gorge. Proceeding upstream about one-quarter mile one comes to a portion of the valley where relations are very clearly shown (see map). Both walls of the old gorge are revealed and the bottom is seen along the southwestern edge and in two places near the middle of the gorge. Proceeding still farther upstream along the northeast bank, one may follow a small stream which enters the amphitheater from the east. For a few hundred feet the course is along a narrow valley excavated in drift. Then, suddenly, one comes upon a nearly perpendicular rock wall which the stream has encountered in its downcutting and over which it falls in a cascade. The rock wall is straight and disappears at both ends under drift. Following southeastward along the projected line of this rock wall, one crosses a hill of drift and soon comes down to the continuation of the wall on the other side (*B*, Fig. 2). Here it is fully exposed, perpendicular, and about 60 feet in height. At the northern end of the exposure the gorge wall is glaciated. The perpendicular cliff is smoothed and striated in true glacial fashion (Fig. 5). From here the gorge wall may be traced almost continuously upstream to the postglacial gorge just below the reservoir, where, after being crossed by the main stream, it disappears under drift (near *W*).

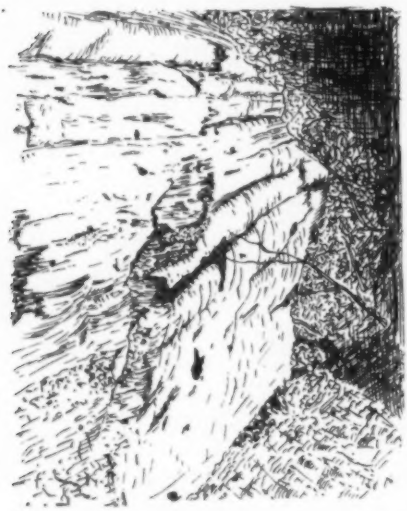

FIG. 5.—The glaciated rock wall of the 600-foot gorge disappearing under drift. From a photograph.

The southwestern wall may be traced with practical continuity all the way. It is, in general, less steep than the northeastern wall.

A steep slope in the place of a vertical cliff is the rule. This southwestern wall also disappears under drift just below the reservoir.

The large amphitheater (just outside the limits of Fig. 2) in which the reservoir lies represents the old gorge again occupied by the stream and locally widened to an unusual extent. Here the gorge walls on the south are steep, but not vertical. On the northeast side the wall appears in several places where small side streams have cut through the overlying drift. It is generally a perpendicular cliff.

The bottom of the old 600-foot gorge, as exposed in a number of places, rises upstream at a rate slightly less than that of the present stream. As a result, the bottom, which appears some 6 feet above the stream at the pumping plant, reaches stream level at the reservoir and lies below it farther up. Partly for this reason the old gorge, in the upper part, becomes more and more indistinct. Its general width, taking the mile of its course just described, is from 500 to 700 feet at the bottom and perhaps 100 feet more than this at the top, for the walls are not everywhere perpendicular. Its relation to the broader glacial trough of Six Mile Creek valley is well shown in the accompanying cross-section (Fig. 3), drawn to a vertical and horizontal scale of 2,000 feet to the inch. The profile makes evident the broad, U-shaped valley with the sharply cut gorge in its bottom. This gorge was probably nearly filled by drift during the various glacial epochs since its formation. In two cases—at the reservoir and south of Green Tree Falls—it is still drift-filled to the top; elsewhere the stream has cleared out a good share of the drift.

In two places on the gorge walls glacial striae have been found; one has already been mentioned, the other is at the eastern edge of the pond, just above the pumping plant. The striations show clearly that the gorge has been covered by ice since its formation, but it seems equally clear from the sharpness of the old gorge walls in most places that the action of this ice was weak and did not materially modify the form of the gorge except at the lower end and where the exposure was particularly favorable. It would seem, therefore, that the gorge must have been cut subsequent to the formation of the broad U-trough of Six Mile Creek valley, in the bottom of which the gorge lies. Otherwise it probably would have

been obliterated by the deep glacial erosion which produced the U-trough.

The second gorge is exposed only below the pumping plant. It is now occupied by the stream all the way from a point a few hundred feet below the pumping plant to the Cayuga Lake delta. Its bottom is nowhere to be seen, since it lies below the level of the present stream. Whether or not it goes below the present base level of Cayuga Lake has not been determined. The fact that the stream is not flowing on a rock bed is not necessarily proof of its having been cut to a lower base level, for it may be that we fail to find rock in the bed of the stream because of the aggradation of the stream bed in connection with the formation of the alluvial fan on which the city of Ithaca stands.

At the Stewart Avenue bridge the gorge is 200 feet wide at the top and has nearly perpendicular walls. Traced upstream, it disappears under drift on the south side of the rock island previously described (*A* of Fig. 2), while the stream lies to the northeast in a postglacial gorge. At the point of disappearance the old gorge has a measured width, from rock to rock of the valley sides, of 125 feet at the bottom. It is nowhere seen farther upstream. It is to be presumed that it lies buried somewhere within the bottom of the older gorge. If so, it must pass under the eastern arm of the pond, for rock exposures preclude its presence elsewhere.

The gorges of the third series are postglacial. They occur where the stream, in cutting down through the later deposits of drift, happened to find itself outside its former gorge. In Six Mile Creek there are three postglacial gorge sections—one just below the pumping plant, another just below the reservoir, and the third just above Green Tree Falls. This latter is being utilized as the site of a second dam for a reservoir for the Ithaca water supply. In the case of the lower of these, at the pumping plant, the diversion of the stream was probably due not so much to filling of the older gorge by glacial till as to the delta deposits built by the stream into Cayuga Lake when it stood at one of its higher levels. A section at the upper end of the rock island by the pumping plant shows till at the base overlain by finely laminated lake clays about 4 feet in thickness, on which lies an undetermined thickness of delta

gravel. This delta deposit is well shown in a cut on the east side of the road, where it was used for gravel. It lies at the general level of the lowest large deltas of Coy Glen and Butternut Creek. There is some evidence, not yet fully worked out, which points to similar deposits as the cause of the diversion of the stream into the two other postglacial gorges farther upstream.

The postglacial gorges are narrow, with walls for the most part vertical rock cliffs. The lower and upper of these gorges still possess waterfalls and cascades. In every case they are distinctly smaller than even the second of the older gorges.

INTERPRETATION

The interpretation put upon the series of gorges in Six Mile Creek is that they were formed by the stream during interglacial intervals. The sequence of events is interpreted as follows:

1. *Preglacial time.*—The stream probably flowed in a broad, mature valley, tributary to the Cayuga Lake trough. The bottom of this valley may have lain at the level of the pronounced rock bench which is clearly marked along the south side of the valley between the 800- and the 1,000-foot contours. This bench is well shown on the topographic map, and is brought out clearly in the accompanying profile (Fig. 3).

2. *First glacial epoch.*—The Cayuga trough, extending parallel with the direction of ice movement, was greatly deepened by ice erosion while at the same time Six Mile Creek valley was deepened considerably and given the flaring U-form typical of glacial erosion. The deepening of the Cayuga trough exceeded that of the Six Mile Creek trough, leaving the latter hanging.

3. *First interglacial interval.*—During this interval Six Mile Creek cut a gorge in the bottom of its hanging valley. That gorge is the oldest, or 600-foot, gorge, described above. The interglacial interval must have been long, for the gorge was cut to an even gradient throughout the whole extent now visible and was also widened very considerably. As we have pointed out before, the gradient of the stream that developed during this interglacial interval was flatter than that of the present stream. An attempt was made, by taking several points as definitely located as possible

on the rock bottom of the old gorge, to plot its profile to scale (see Fig. 6). Points were taken at intervals through a distance of over a mile along the gorge. It was found that, when these points were plotted, the line connecting the first and last passed very close to all the others. This would show that the gorge bottom was well graded. When this profile of the gorge is projected out over the Cayuga valley it is found that at the edge of the Cayuga trough the bottom of the gorge must have been hanging 80 feet above the present lake level. Projecting this line still farther to the axis of the trough, we find it 28 feet above the present lake level. This

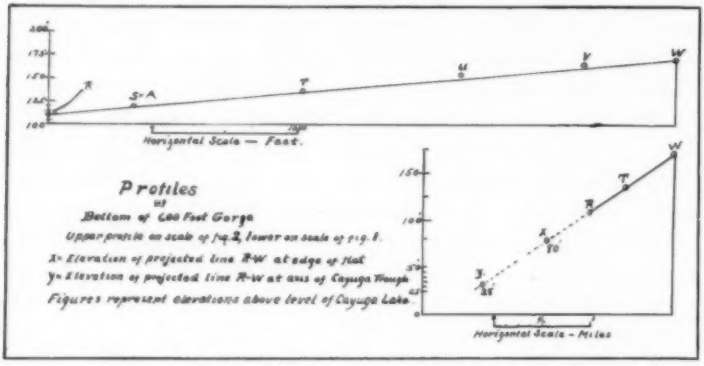
Horizontal Scale — Feet.
Profiles
of
Bottom of 600 Foot Gorge
Upper profile on scale of fig. 2, lower on scale of fig. 1.
X= Elevation of projected line R-W at edge of flat
Y= Elevation of projected line R-W at axis of Cayuga Trough
Figures represent elevations above level of Cayuga Lake
Horizontal Scale - Miles

FIG. 6.—Profiles of the rock bottom of the 600-foot gorge. Elevations, taken from the Ithaca water survey map, represent height above Cayuga Lake level. For height above sea add 381 feet.

would indicate that during the first interglacial interval the base level in the main valley stood approximately 80 feet above the present lake level. This base may have been a stream in the main valley or it may have been a lake similar to the one now occupying the basin. We shall discuss later the evidence bearing on this point.

4. *Second glacial epoch.*—A second glacial epoch followed and so lowered the base level in the main valley that the tributary stream again found itself hanging. The ice, during this glacial invasion, seems to have considerably modified the lower part of the 600-foot gorge. The sides and bottom were smoothed and striated and the walls of the gorge at the lower end appear to have been much eroded

and partly destroyed. These gorge walls become practically unrecognizable below the rock island. Without a doubt some of this erosion was done by the later, Wisconsin, ice sheet, but it is thought that the greater part was earlier. The reasons for this belief will be stated in a later paragraph.

5. *Second interglacial interval.*—At the close of the second glacial epoch Six Mile Creek, again left hanging, began cutting another gorge—the 200-foot gorge through which the stream now flows from the rock island to the alluvial flats of the Cayuga valley. The base level of this gorge must have been as low as the present lake level, for rock is nowhere exposed in the stream bed. How much below the present lake level the intergiacial base may have been we have no means of knowing. The time interval represented by this gorge must have been much shorter than that during which the older, 600-foot, gorge was cut, for, as has been stated before, the width of the gorge is not much more than one-third that of the older gorge. Judging from the size of the gorge, however, the time must have been much longer than the interval since the last glacial epoch, for the gorge was cut back at an approximately even grade for a distance of two-thirds of a mile at least—how much farther we have no means of knowing, for it is buried under drift. Postglacial streams of equal size have succeeded, at best, in grading their gorges only a few hundred feet back from the Cayuga valley trough. In fact, most of them come tumbling down over the rocks of the hillsides through gorges which are merely notches in comparison with the interglacial gorges.

6. *The third glacial epoch, Late Wisconsin.*—This ice invasion put an end to the gorge-cutting. It filled or partly filled the older gorge with drift and at its close delta deposits further clogged the gorges until, as the lakes were lowered, the stream found its old channel so completely blocked in places that it must seek lower ground to one side. This determined the stream's course.

The ice of the Wisconsin epoch seems to have done little toward modifying the form of the second gorge. Its walls are still steep and angular and on them we know of no glacial striations having been found. It is the fresh appearance of this second gorge, as compared with the glacially eroded lower end of the 600-foot gorge,

which led us to believe that the Wisconsin ice had played but a small part in the widening and modification of the 600-foot gorge. It seems to be still an open question whether the ice of the *latest* Wisconsin epoch ever extended beyond the belt of strong terminal moraines just south of Ithaca.[1] If it did not, we should expect only feeble erosive action from the Wisconsin ice in the Ithaca region. It is interesting to note, in this connection, that Tarr, in his report on the Watkins Glen-Catatonk Folio, emphasizes the slight erosion by Wisconsin ice.

7. *Postglacial interval.*—This has been marked, in the case of Six Mile Creek valley, by the partial re-excavation of the older gorges and by the cutting of postglacial gorges in places where the postglacial stream found itself outside its former course. The postglacial seems to have been the shortest of the intervals of deglaciation. The gorges are relatively small and show much less stream erosion than the others. In spite of the fact that climatic differences may have influenced the rate of gorge-cutting, it seems legitimate to judge, roughly, the length of an interglacial interval by the size of its gorges.

ALTERNATIVE HYPOTHESES

Before the interpretation outlined above can be considered established, certain questions call for consideration. The first of these is whether the observed relation of the 200-foot gorge to the 600-foot gorge—that is, cut in its bottom—necessitates for its explanation the intervention of a glacial epoch; the second is whether the 600-foot gorge may not be a product of preglacial rejuvenation.

In answer to the first of these questions it may be said that the whole problem resolves itself into a search for an adequate reason why the stream, after cutting the 600-foot gorge to a low gradient—evidently to its local base level—and widening it considerably, should suddenly have renewed its downcutting sufficiently to have formed the 200-foot gorge in the bottom of the older one. Several possible explanations present themselves: (*a*) there may have been a lake in the main valley whose level was suddenly lowered; (*b*) the

[1] Chamberlin, in his report on the "Terminal Moraines of the Second Glacial Epoch" in the *Third Annual Report of the U.S. Geol. Survey*, states, as a possibility, that the moraine in question represents the limit of Wisconsin ice movement. It seems to the writer that this possibility has never been satisfactorily set aside.

main valley may have been occupied by a stream which in some way, perhaps by uplift, was suddenly rejuvenated; (*c*) a glacial invasion may have altered the base level in the main valley. In regard to these possible explanations, we believe it not only possible, but very probable that a lake, similar to the present Cayuga, existed in the valley previous to the last ice invasion. It is conceivable, also, that such a lake might be drained comparatively suddenly. For instance, a hard rock stratum might be so disposed as to form a barrier which would yield rather suddenly and expose weaker rocks beneath. It is evident, therefore, that we cannot at once dispose of hypothesis (*a*). On the other hand, we have in Six Mile Creek alone no means of proving that a river in the main valley may not have been suddenly rejuvenated. It will be apparent, however, that if such a trunk river, by its rejuvenation, led to the formation of a gorge in Six Mile Creek valley, it should, in a similar manner, have led to the formaton of similar gorges, corresponding to the 200-foot gorge of Six Mile Creek, in all the other tributaries. We should expect to find, then, that the wide buried gorge of Buttermilk Creek, which apparently corresponds to our 600-foot gorge, should have a narrower gorge, corresponding to our 200-foot gorge, sunk in its bottom. Such is not the case, however. We must conclude from this, we believe, that hypotheses (*a*) and (*b*) are both untenable, for the argument stated above would apply equally well in the case of a lake whose level was suddenly lowered. The third hypothesis suggested—the lowering of the base level of the tributaries by the intervention of a period of glacial erosion—does not meet with this objection, for it might well happen that the Six Mile stream, after the disappearance of the ice, would find itself still in its old channel, while the Buttermilk stream, happening to find its old channel blocked by drift or delta deposits, cut a new gorge elsewhere, as, in fact, it has done.

A careful study of the buried gorges of Butternut Creek, Newfield Creek, and others should be made in order to settle the question beyond possibility of dispute.[1]

[1] In this connection it may not be out of place to call the attention of any whose lot it may be to make a further study of these buried gorges to a remarkably fine buried

To return to a consideration of the second question, namely, whether the 600-foot gorge may not be a product of preglacial rejuvenation, we have seen that lower Six Mile Creek valley was profoundly deepened and modified by ice erosion and that it was given the broad, flaring U-form of a typical glacial trough. The 600-foot gorge now lies in the bottom of this trough and still retains perpendicular or nearly perpendicular rock walls throughout much of its extent. If this gorge was formed by a preglacial stream it must have survived the deep scouring which lowered the bottom of the valley to an extent of some 200 feet or more and modified its form to that of a typical U-trough. It is hard to believe that such a gorge would not have been widened by the ice and made a part of the U-trough had it been in existence while the erosion was in progress. At least the sharpness of the gorge walls would have been destroyed. It is true that, on the southwest wall of the gorge, this sharpness has been to some extent destroyed, but this is probably the result of erosion by the ice of some of the later periods. The northeast wall is still sharp.

A further point which may have a bearing on this problem is the fact that in the valley of Fall Creek there are drift-filled gorges, as, for instance, those just north of the Triphammer Bridge, whose bottoms are nearly 200 feet above that of the 600-foot gorge in the Six Mile Creek valley. Certainly these gorges cannot belong to the same cycle as the 600-foot gorge, for both seem well graded, yet their base levels are so different. If either is preglacial it must be the higher one. Buried gorges at levels higher than our 600-foot gorge are to be found in several of the valleys of the region besides that of Fall Creek. On detailed study, with careful leveling, it may appear that even the 600-foot gorge of Six Mile Creek belongs to a

gorge in the valley of a small stream which enters Cayuga Lake from the east at Shurger Point. In following up the gorge of this stream one soon encounters a series of cascades and falls aggregating perhaps 75 feet in height. Above these cascades is a gorge 100 feet, more or less, in width at the bottom, with rock walls, and a remarkably even, low-gradient rock floor. The gorge, with these characteristics, continues upstream for half a mile to a point just below the trolley track (one mile from the lake), where it suddenly ends, the rock walls disappearing under drift, and the stream tumbling in over a fine fall 60–80 feet in height. The bottom of the upper section of this gorge has a low gradient and hangs at a level somewhere between 60 and 100 feet above that of Cayuga Lake.

comparatively late interglacial interval, and that some of these higher gorges represent still earlier intervals, thus still further increasing the complexity of the glacial record.

LITERATURE

The interpretation of these gorges which we have presented is not, in all respects, new. A brief summary of the literature pertaining to Six Mile Creek and its gorges will indicate the status of the problem previous to the recent investigations.

Probably the first mention of the drift-filled gorges was made by Simonds in 1877,[1] who says: "The valley of Six Mile Creek furnishes some special examples of the drift phenomena. In several places its old channel has been completely choked up with masses of morainic débris, about which the present stream has been obliged to cut its way through deep canyons." This reference is interesting, though it contains nothing bearing directly on the problem in hand.

In the *Physical Geography of New York State* (New York: Macmillan Co.), published in 1902, Tarr, under the heading "Interglacial(?) Gorges" (pp. 178–79), calls attention to "numerous gorges which are broader than the postglacial valleys and partially obscured by glacial till, showing that they were formed either during preglacial or interglacial times," and mentions Six Mile Creek as an especially good example. The problem which has ever since been recurring, and, in fact, still calls for discussion, is there clearly stated: "Were the gorges [of central and western New York] due to the interglacial conditions or to an uplift in preglacial times?"

In a paper on "Hanging Valleys in the Finger Lake Region of Central New York"[2] Tarr discusses the Six Mile Creek gorges at some length. He points out the fact that practically all the valleys tributary to the Finger Lakes possess gorges "which antedate the last advance of the ice" and that in the case of Six Mile Creek

the stream alternately enters the buried gorge, forming broad amphitheaters, and where it for a short time leaves the earlier gorge, crosses spurs of rock in

[1] *American Naturalist*, XI (1877), 49–51.

[2] *American Geologist*, XXXIII (May, 1904), 271–91.

narrow, postglacial gorges. . . . . A third condition is where a stream enters the earlier gorge, clears out part or all of its drift filling, and follows it across the steepened slope to the main valley. This is certainly the case in lower Six Mile Creek.

It should be noted that in this reference to Six Mile Creek Tarr makes no distinction between the gorge characterized by the "broad amphitheaters" mentioned above (our 600-foot gorge) and the gorge of "lower Six Mile Creek," which the stream follows to Cayuga valley (our second, or 200-foot, gorge).

A following paragraph (p. 281) is especially significant in connection with the facts brought out by our recent study. We quote in full:

This evidence establishes a third point regarding the drainage history of Seneca and Cayuga valleys, namely, that there is an almost, if not absolutely, uniform condition of gorges in the bottoms of the mature hanging tributary valleys; that these gorges, being drift-filled, antedate the last ice advance; that they are broader and deeper, hence required longer time to form than the postglacial gorges; and that, where they enter the main valleys, their rock bottoms are above lake level. They therefore resemble hanging valleys, since their bottoms are in some cases, as Taughannock, about 400 feet above the main valley bottom one-half mile distant. Since we have no data proving what their bottom slope actually is, though it seems evident that their slope is very steep, and possibly great enough to carry them down to the main valley axis, it may not be proper to consider them hanging valleys. If the interpretation of hanging valleys is warranted, which I doubt, then the tributary valleys to Seneca and Cayuga Lakes are double hanging valleys—an upper mature hanging valley and a lower hanging gorge valley.

In the case of Six Mile Creek at least we now have the data proving the actual bottom slope (Fig. 6) and it appears that the slope is, in reality, comparatively gentle, and that the double hanging valleys of which Tarr speaks are, in fact, present.

In another paragraph in the same paper (p. 284) the results of our recent study are foreshadowed in the following words:

A modification of the glacial erosion theory has been advanced during the progress of the investigation of the problem and is still being considered. It is as follows: During its first advance the ice deeply eroded the valleys; during interglacial conditions the older valleys were cut; with return of glaciation the valleys were deepened still further. During as many glaciations as this region experienced this process was continued. On this basis the older gorges are interglacial; their cause is the lowering of their base level by the overdeepening of the valleys to which they were tributary. Since facts sufficient to establish

or to overthrow this explanation are not yet at hand it must stand at present merely as a working hypothesis.

The question of the explanation of the buried gorges by regional rejuvenation is then raised (p. 290). Two sentences in connection with this are quoted:

> Distinct progress toward solution might be made if it were possible to establish the grades of the buried gorges. If they can be shown to have such steep grades as to carry them down to the main valley bottoms the theory of rejuvenation will be greatly strengthened.

In this connection we again call attention to Fig. 6, showing the projected grade of the rock bottom of the oldest of the buried gorges of Six Mile Creek, which profile, it would seem, does not tend to strengthen the rejuvenation theory.

As late as 1906 we find the following in regard to the gorges of the Finger Lake valleys:[1]

> There are some facts which indicate possible greater complexity of ice erosion, for in some of the valleys there is apparently more than one buried gorge; but the evidence on this point is not as yet convincing, and for the present we can point with certainty to no greater complexity than that of two periods, one the Wisconsin, the other of some one of the earlier ice advances, with which the work of the glacial geologists of the Mississippi Valley have made us familiar.

Tarr's latest published conclusions in regard to these buried gorges may be found in the Watkins Glen-Catatonk Folio of the United States Geological Survey, 1909. He concludes that evidence of a pre-Wisconsin period of ice erosion "is afforded by the presence of hanging gorges, partly buried in Wisconsin deposits. . . . . These are evidently interglacial gorges cut in the bottoms of hanging valleys that were left hanging by overdeepening of the main troughs through ice scouring" (p. 117, field edition); and again, p. 224: "We have conclusive evidence here [referring to the buried gorges] of only two advances, but there may have been more."

Matson (*op. cit.*) at an earlier date had described buried gorges in the valley of Buttermilk Creek, near Ithaca, which led him to the conclusion that in that valley there exists a "series of complex gorges which are considered interglacial. The minimum number of epochs of deglaciation is two; the maximum number, four. . . . ."

[1] R. S. Tarr, "Watkins Glen and Other Gorges of the Finger Lake Region of Central New York," *Pop. Sci. Mo.* (May, 1906), 387–97.

It will be evident from the foregoing summary of the literature that more than one set of buried gorges has been recognized, or at least suspected. The problems still left unsolved were: (1) the details of these gorges, and the number present in different valleys; (2) the gradient of the gorge bottoms; (3) the possibility of the explanation of one of the buried gorges by preglacial rejuvenation.

BROADER BEARINGS OF THE PROBLEM

The buried gorge phenomena of Six Mile Creek do not stand isolated. Similar phenomena in many other valleys of the region have been recorded.

If our interpretation is correct they all form a definite part of the sequence of events connected with the glacial epochs. Some valleys should show several gorges; some, perhaps, in cases where the stream happened to find the same channel each time, should show only one. The phenomena in Six Mile Creek valley point to at least three glacial epochs separated by periods of time longer than that since the last epoch. Other valleys may show an even more complete record. Buttermilk valley shows a more complex series than Six Mile, and Butternut Creek has a splendid series of old gorges, very little studied as yet.

As the study of these gorges progresses it should be possible to settle many points now in question. For instance: the 600-foot gorge of Six Mile Creek was cut to a base level approximately 80 feet above present lake level. The largest gorge of Buttermilk was cut to approximately the same base. The base of the largest gorge in Butternut Creek lies at approximately the same level—perhaps somewhat higher. By careful leveling it should be possible to determine with close approximation the base level to which each of these gorges was cut. It may be possible in this way to determine whether a lake (horizontal base level) or a river (inclined base level) occupied the main valley at the time of the cutting of the gorges. The Finger Lake valleys and their tributaries afford abundant evidence which must be studied carefully before a final decision as to the full complexity of the glacial period in central New York can be made.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA FOR 1909, 1910, 1911, AND PART OF 1912

EDWARD STEIDTMANN
University of Wisconsin

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I. General
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1910
1911
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IV. Ontario, Northwest Territories, and East Coast of Hudson Bay
V. Quebec
VI. The Cordilleras

I. GENERAL

1909

Adams[1] maintains that the pre-Cambrian may be divided on the basis of major diastrophic movements into Eo, Meso, and Neo-Proterozoic, and that this basis may serve the purpose of international correlation better than comparative study of unconformities. He objects to the two fold classification of the pre-Cambrian suggested by Van Hise, on the ground that the break at the base of the Animikie is as great as that which follows the Keewatin.

Adams[2] states that amphibolites consisting of aggregates of hornblende, pyroxene, mica, and other minerals, in various proportions, have developed from (*a*) the metamorphism and recrystallization of impure calcareous sediments; (*b*) alteration of certain basic dikes; and (*c*) alteration of limestones by batholithic intrusions of granite. The granites show marginal digestion of limestones

[1] F. D. Adams, "The Basis of pre-Cambrian Correlation," *Jour. Geol.*, XVII, No. 2 (1909), 105–23.

[2] F. D. Adams, "Origin of Amphibolites of the Laurentian Area of Canada," *Jour. Geol.*, XVII, No. 1 (1909), 1–18.

grading into a mosaic of hornblende and feldspar crystals. With diminishing hornblende the gradation passes into a pyroxene, scapolite, biotite rock, and finally into coarse crystalline limestone.

From analyses of limestones representing three stages of alteration to amphibolite near a contact, Adams concludes that the alteration has involved the introduction from the granite of silica, alumina, iron, magnesia, and alkalies, and the loss of lime and carbonic acid. This conclusion seems to rest on the assumption that the alteration took place without change of mass.

Lindgren[1] presents a résumé on the principal epochs of the segregation of metals on the North American continent. Many different epochs of ore formation are embraced in the pre-Cambrian period (in many different places). In the eastern part of the continent iron, copper, nickel, silver, and gold ores occur in association with a variety of both intrusive and extrusive igneous rocks ranging from granite to basalt. In the Cordilleran region, intrusives, principally granites, are almost exclusively represented. Intrusive diorite, gabbro, and diabase are present. The Cordilleran pre-Cambrian ores are principally gold and silver. The lead and zinc ores of the pre-Cambrian are unimportant.

Van Hise[2] maintains that the bases of classification of the pre-Cambrian should be physical, those of most importance being: (1) lithologic character, (2) continuity of formations, (3) likeness of formations, (4) like sequence of formations, (5) subaerial or subaqueous deposits, (6) unconformities, (7) relations to series of known age, (8) relations with intrusive rocks, (9) amount of deformation, (10) degree of metamorphism. His major divisions of the pre-Cambrian are Archaean and Algonkian.

1910

Adams[3] concludes that large igneous intrusives rising from the deeper portions of the earth are one of the principal agents of the

[1] Waldemar Lindgren, "Metallogenetic Epochs," *Econ. Geol.*, IV, No. 5 (1909), 409–420.

[2] C. R. Van Hise, "Principles of Classification and Correlation of the pre-Cambrian Rocks," *Jour. Geol.*, XVII, No. 2 (1909), 97–104.

[3] Frank D. Adams, "The Origin of the Deep-seated Metamorphism of the pre-Cambrian Crystalline Schists," *Compte Rendu*, Congrès Géologique International, 1910, pp. 563–72.

metamorphism of rocks, since more and more cases of intimate relationship between the occurrence of schists and batholithic intrusions, principally granite, have been found as geologic study proceeds.

Coleman[1] presents a summary of the stratigraphy and correlation of the pre-Cambrian as adopted by the international committees on correlation, before the British Association at Winnipeg.

Coleman[2] argues that the abundance of carbonaceous shales, limestones, and other ordinary sediments in the Keewatin of Ontario implies the existence of life and climatic conditions not unlike the present.

Coleman[3] infers that uniformitarian conditions prevailed in the pre-Cambrian from the similarity of pre-Cambrian sediments to those of later periods.

Coleman[4] presents a summary on the degree and kind of metamorphism of the various pre-Cambrian formations of northern Ontario. The agents of metamorphism as outlined by Coleman are heat, pressure, and solutions.

Coleman[5] presents a historic summary of the methods which have been employed in classifying the Archaean of Ontario. The principal methods, according to Coleman, in the order of their historic development are relative stratigraphic position of formations, lithologic similarity, unconformity, eruptive contact, and basal conglomerates.

Coleman[6] summarizes the principal known events in the history of the Canadian shield.

[1] A. P. Coleman, "The pre-Cambrian Rocks of Canada," *British Assoc. Adv. Sci. Rept.* 79th Meeting, 1910, pp. 474–75.

[2] A. P. Coleman, "Climate and Physical Conditions of the Keewatin," *Geol. Soc. Am. Bull.*, XXI, No. 4 (1910), 778–79.

[3] A. P. Coleman, "The Bearing of pre-Cambrian Geology on Uniformitarianism," *British Assoc. Adv. Sci. Rept.*, 79th Meeting, 1910, pp. 473–74.

[4] A. P. Coleman, "Metamorphism in the pre-Cambrian of Northern Ontario," *Compte Rendu*, Congrès Géologique International, 1910, pp. 607–16.

[5] A. P. Coleman, "Methods of Classification of the Archaean of Ontario." *Compte Rendu*, Congrès Géologique International, 1910, Vol. I, pp. 721–33.

[6] A. P. Coleman, "The History of the Canadian Shield," *Nature*, LXXXIV (September 15, 1910), 333–39.

R. A. Daly[1] believes that the pre-Cambrian ocean probably contained less lime than the present ocean, for the following reasons: the sea must have been fresh or nearly fresh initially; the pre-Cambrian rivers probably carried less lime because the lands of that time did not possess the great development of limestone of the present; the lime contributed to the sea was probably precipitated by ammonium carbonate generated through the decay of organisms which accumulated on the sea bottom because of the lack of scavengers. The field evidence which he cites in favor of this hypothesis is the fineness of grain of many undisturbed pre-Cambrian dolomites which he believes is indicative of chemical precipitation, and the general scarcity of fossils in pre-Cambrian rocks, which he ascribes to the lack of lime in the sea from which organisms could obtain the materials for shells and hard parts.

Evans[2] believes that the sudden appearance of the Cambrian fauna may have been due to causes very similar to those which gave rise to the sudden appearance of the Tertiary fauna. These periods were preceded by great uplift of the lands, vulcanism, and widespread aridity. The record of the evolution of life immediately preceding them is therefore largely lost or inaccessible.

Harder[3] states that the pre-Cambrian manganese ores of the United States are found in association with the hematite deposits of the Lake Superior region; as manganiferous zinc ores at Franklin Furnace, N.J.; and in the metamorphic rocks of the Piedmont region of Virginia, North Carolina, South Carolina, and Georgia.

Hayes[4] states that the pre-Cambrian iron ores of the Lake Superior region furnish about 80 per cent of the annual ore production of the United States.

The iron formations from which the ores were derived are sedimentary deposits, consisting mainly of cherty iron carbonates,

[1] R. A. Daly, "Some Chemical Conditions in the pre-Cambrian Ocean," *Compte Rendu*, XI. Congrès Géologique International, 1910, pp. 503–9.

[2] John W. Evans, "The Sudden Appearance of the Cambrian Fauna," *Compte Rendu*, XI. Congrès Géologique International, pp. 543–46.

[3] Edmund Cecil Harder, "Manganese Deposits of the United States with Sections on Foreign Deposits, Chemistry, and Uses," *Bull. 427, U.S. Geological Survey*, 1910, pp. 298.

[4] C. W. Hayes, "Iron Ores of the United States," *Bull. 394, U.S. Geological Survey*, 1910, pp. 76–84.

ferruginous cherts, and ferrous silicate rocks interbedded with clastics and extrusives. Their average iron content ranges from about 25 to 35 per cent iron. The ores were developed by the solution and removal of silica, and the oxidation, solution, transportation, and redeposition of the iron, in the exposed portions of the formations, having structural peculiarities favorable to the circulation of water, such as jointing, pitching basements, faults, etc.

The ores mined are mainly hematite and limonite, although some magnetite has been developed through deep-seated conditions of high pressure and temperature. Their iron content, while steadily falling, has generally been above 50 per cent, but 40 per cent ores are now mined as mixers for higher-grade materials. The estimated tonnage of the available ores, including all ores above 55 per cent iron, and 25 per cent of all formations containing from 45 to 55 per cent iron is 3,500,000,000 tons. Iron formations containing from 35 to 45 per cent iron are classed as available in the future. Their tonnage is enormous, 72,000,000,000 tons being taken as an arbitrary estimate.

The pre-Cambrian Adirondack area of New York contains commercial deposits of non-titaniferous magnetites, titaniferous magnetites, and red hematites.

The non-titaniferous magnetites are intimately associated with sedimentary rocks, marble, schists, and gneisses. Separation is generally made by magnetic methods and the ore shipped to the iron centers, excepting a small tonnage which is smelted in local coke ovens. The ores are both Bessemer and non-Bessemer. The lowest grades contain about 35 per cent iron when mined, the richest from 60 to 65 per cent. The concentrates run from 60 to 65 per cent. No fair estimate of the tonnage can be made. The known deposits of high grade are believed to contain 35,000,000 tons, and the leaner deposits carrying over 35 per cent iron may yield 75,000,000 tons of concentrates.

The ores are believed to have been developed in connection with the intrusion of igneous rocks.

The titaniferous magnetites carry at least 8 per cent of titanic acid and on an average about 15 per cent. They are found at the margins of a gabbro-anorthosite area, and are spoken of as magmatic

segregations. The richest ores contain about 60 per cent iron. They are not used at present because of their titanium content, excepting on a very small scale. Their extent is unknown. In the Lake Sanford district alone, 90,000,000 tons are known.

Karl L. Henning[1] reviewed the principal results of Van Hise and Leith's *Pre-Cambrian Geology of North America.*

Kemp[2] states that the following tentative chronologic table of the Swedish pre-Cambrian based on the work of Sederholm and Högbom is finding favor:

Upper or Jotnian	Epijotnian dislocations
	Jotnian
	Subjotnian land surface denudation and igneous rocks
Middle or Jatulian	Epijatulian folding
	Jatulian
	Subjatulian land surface and denudation
Lower or Archaean	Serarchaean granites
	Archaean
	No chronological subdivision. Difference due to varying metamorphism

Kemp shows that the Archaean of Sweden is separable into two major divisions, an older complex of minor sediments and major deep-seated intrusives, and a younger series of intrusive granites consisting of four distinct types. Similarly, the Archaean of the Lake Superior region embraces an older division consisting of minor sediments and major igneous rocks, mostly basic extrusives, however, and a younger division consisting almost entirely of granitic intrusives. The sediments of the Swedish Archaean contain squeezed conglomerates, crystalline limestones and dolomites, mica schists, leptites, garnet, cordierite, sillimanite, and graphite gneisses of probable sedimentary origin. The leptites are fine-grained, banded, and stratified rocks, which include sediments and iron ores. The Archaean sediments of the older division in the Lake Superior region include iron formation and minor clastics. The Archaean igneous rocks of Sweden of the older division comprise granites, gneisses, and gneissoid intrusives, titaniferous ore-bearing gabbros,

[1] Karl L. Henning, "Die pro-kambrische Geologie von Nord Amerika," *Naturwissenschaftliche Wochenschrift*, N.F., Bd. No. 28 (July, 1910), pp. 433–37.

[2] J. F. Kemp, "Comparative Sketch of the pre-Cambrian Geology of Sweden and New York," *New York State Museum Bull. 149*, 1910, pp. 93–119.

syenites, and anorthosites. The leptites, sediments, and iron ores are compared by Kemp to channels amid an archipelago of large island intrusives.

The surface on which the Jatulian was deposited is regarded as hilly by Högbom and mountainous by Törnebohm. Kemp believes that the time gap of erosion which separated the Jatulian from the Archaean was not so long as the one which separated the Cambrian from the Jotnian or the Jotnian from the Jatulian. The Jatulian in Sweden consists of quartzite, schists, dolomitic limestones, and beds of anthracite, which have a maximum thickness of two meters.

Professor Sederholm has divided the Jatulian of Finland into an upper and lower member which consist of eruptives, greenstones, clastics, and dolomites, and has compared it to the Upper Huronian of the Lake Superior region. The Jatulian closed with a period of very intense folding, followed by intrusions of a peculiar porphyritic granite which decomposes readily, forming the so-called *rapakivi*, Finnish for rotten stone. This has afforded an important guide by which Swedish geologists have measured the age of various other intrusives and metamorphic rocks. The erosion which followed developed an even floor free from weathering products, on which the Jotnian diabases and ripple-marked and sun-cracked sandstones were deposited. These now lie in isolated patches. They are compared by Professor Kemp to the Torridonian of Scotland and the Keweenawan of the Lake Superior region. Finally, Cambro-Silurian sediments of which widespread relicts remain were deposited upon a generally even surface covered with a weathering breccia which grades downward into a kaolinized gneiss. So-called sandstone dikes containing Cambrian fossils are found in gneisses far from Cambrian beds.

Most of the comparisons made by Professor Kemp between the pre-Cambric of Sweden and New York are necessarily petrographic and appeal to those who are personally familiar with the rocks of the New York pre-Cambrian.

Lane[1] finds that the grain of Laurentian granites tends to be uniform from the margin to the center, from which he infers that

[1] A. C. Lane, "The Stratigraphic Value of the Laurentian," *Compte Rendu*, XI. Congrès Géologique International, 1910, pp. 633–37.

the country rock which the granites invaded must have been nearer the condition of crystallization or fusion than the granites, and that the Laurentian granites are batholiths rather than re-fused sediments or portions of the original crust of the earth.

Lindgren[1] states that the principal pre-Cambrian ores of North America are iron, copper, nickel, gold, and silver. The iron ore types are: magnetite and ilmenite of igneous origin, and hematites of sedimentary and igneous origin, which have been concentrated by surface waters. The copper, nickel, and silver of the Lake Superior region and the copper ores of the South are connected with basic igneous rock. The gold quartz veins of the south are related to granitic intrusions.

Newland[2] states that the iron ores mined in Sweden are all pre-Cambrian magnetites associated with crystalline schists and acid igneous rocks. There are also some low-grade, gabbroic, titaniferous magnetite silicate rocks, and bog lake deposits, the latter famous as exemplifications of present-day ore deposition, but these are not mined.

Most of the ore comes from central Sweden. A considerable portion of this is unique for its low phosphorus content, and because of its exceptional qualities can be mined from very small deposits on a miniature scale. The ore is reduced in charcoal furnaces. Nearly half of the output of central Sweden is high phosphorus ore from Grangesberg. The high phosphorus ores are all exported, since there are no coals in Sweden suitable for furnaces. The ores of central Sweden occur in association with sediments, leptites, and gneisses which form winding lenses between massive intrusives. The ores at Norberg and Striberg consist of banded magnetite quartz rocks with a sedimentary aspect.

The Kiruna and Gellivare magnetite apatite deposits of Lapland resemble the titaniferous ores of New York in their association with sodic rocks. The ores at Gellivare are lenses, bands, and chimneys in syenite, the entire mass showing regional metamorphism. The

[1] W. Lindgren, "Metallogenetic Epochs," *Jour. Can. Min. Inst.*, XII (1910), 102–13.

[2] D. H. Newland, "Notes on the Geology of the Swedish Magnetites," *New York State Museum Bull. 149*, 1910, pp. 107–19.

ores at Kiruna are of the nature of large dikes between igneous rocks. The largest, that of Kurunavaara, is 5 kilometers long and from 50 to 164 meters thick, and lies between a quartz porphyry hanging wall and syenite porphyry foot. The ore has a dense, steely appearance, and consists of magnetite with a little apatite. They are probably igneous in origin because of the igneous character of the wall rocks, the igneous texture of the ore, the presence of apatite dikes in the ore and wall rock, and the existence of apophyses of ore in the footwall and the surrounding porphyry.

Sederholm[1] believes that certain pre-Cambrian graphitic schists and limestones indicate the existence of progonozoique forms. The progonozoique formations alluded to are the formations which contain vestiges of the ancestral forms of the life which suddenly appears in the Cambrian.

Sollas[2] ascribes the scarcity of fossils in the pre-Cambrian to a feeble development of hard parts in the organisms of that period. From the rough approximation that "the ontogeny of the individual repeats the phylogeny of the race," he infers that the organisms of the pre-Cambrian were similar to the larval forms of the existing invertebrata. Such larval forms are not known to have been preserved, notwithstanding that some are endowed with calcareous skeletons.

Walcott[3] accounts for the sudden appearance of abundant marine life with the Cambrian by assuming that marine life first developed in the open ocean and did not migrate to the shore zones until the Algonkian uplift preceding the Cambrian had largely excluded the seas from the continents.

Walther[4] ascribes the scarcity of fossils in certain pre-Cambrian formations immediately below the Cambrian to their deposition

[1] J. J. Sederholm, "Sur les vestiges de la vie dans les formations progonozoique," *Compte Rendu*, XI. Congrès Géologique International, 1910, pp. 515–24.

[2] W. J. Sollas, "The Fauna of the Protaeon," *Compte Rendu*, XI. Congrès Géologique International, 1910, Vol. I, pp. 499–501.

[3] C. D. Walcott, "Abrupt Appearance of the Cambrian Fauna on the North American Continent," *Smithsonian Misc. Coll.*, LVII, No. 1 (1910), pp. 1–16, 1 pl., 1 fig.

[4] T. Walther, "Die lithologischen Eigenschaften der Gesteine im Liegenden der kambrischen Formations," *Compte Rendu*, XI. Congrès Géologique International, 1910, Vol. I, pp. 511–513.

under desert conditions. He applies this hypothesis to the Algonkian of the Grand Canyon, the Torridonian sandstone of Scotland, and the Sparagmite formation of northern Scandinavia.

Matthew[1] ascribes the sudden appearance of Cambrian life to the obliteration of fossils by chemical changes and other causes, partly to the absence or rarity of Benthos in the earlier faunas, and partly to the fact that fossiliferous strata beneath the Cambrian are arbitrarily attached to the base of the Cambrian.

1911

C. K. Leith and E. C. Harder[2] state that Brazilian hematite ores of Minas Geraes are found about 300 miles from the coast, in a region constituted by a basement complex of crystalline schists, upon which rests a sedimentary series consisting of slates, schists, quartzites, ferruginous quartzites or itabirites with interbedded hematite ores, gneiss, ferruginous carbonate rocks, carbonate rocks, and amphibolites. The sediments have been severely folded and altered by granitic intrusives. Near the bottom quartzite predominates. Argillaceous sediments predominate near the top and the iron ore beds are most abundant near the middle of the sedimentary series.

The most abundant ores of the district are thin-bedded, fine-grained hematites, of excellent grade. Thick-bedded, massive, nearly pure hematites rank next in importance.

A very large tonnage of fragmental ore consisting of residual hematite and more or less transported débris resulting from the breakdown of the ore deposits is in sight. These ores are cemented with sands, clays, and limonite and therefore are lower in grade than bedded deposits. A negligible quantity of ore is known to have developed from the leaching of ferruginous carbonates and itabirite.

The Lake Superior ores have resulted from primary deposition and secondary concentration. In no known case has primary deposition alone developed ore. All the ores are related to structu-

[1] G. F. Matthew, "The Sudden Appearance of the Cambrian Fauna," *Compte Rendu*, Congrès Géologique International, 1910, pp. 547–551.

[2] C. K. Leith and E. C. Harder, "Hematite Ores of Brazil and a Comparison with Hematite Ores of Lake Superior," *Econ. Geol.*, VI (1911), 670–86.

ral features which favored the circulation of water. In Brazil, the iron ores are the result of primary deposition, and while some secondary concentration by leaching has taken place, it is local and very subordinate. The fragmental ores of Brazil have no counterpart in the Lake Superior region excepting in conglomerates at the base of formations resting on ore deposits. Glacial erosion may have removed large quantities of fragmental ore in the Lake Superior region.

Miller[1] points out that certain pre-Cambrian stratigraphic units of the Northwest Highlands of Scotland are similar to some of Canada as indicated by the following table:

Canada	Scotland
A. Keweenawan.	A. Torridonian.
B. Huronian with intrusives.	B. Intrusives of Lewisan. A fragmental series of quartzite, etc., has been removed before Torridonian was deposited.
C. Keewatin-Laurentian Complex. Granite and gneiss, greenstone, limestone, iron formation, etc.	C. Lewisan. Fundamental Complex. Gneiss, hornblende and chloritic schists, limestone, and iron formation.

Miller and Knight[2] believe that the present use of the term Laurentian as applied to the granites and granite gneisses intrusive into the Keewatin but presumably not in the Lower Huronian may have led to poorly based correlations.

[1] W. G. Miller, "A Geological Trip in Scotland," *Ont. Bur. Mines, 20th Ann. Rept.*, 1911, pp. 259–69.

[2] W. G. Miller and C. W. Knight, "The Laurentian System," *Ont. Bur. of Mines, 20th Ann. Rept.*, 1911, pp. 280–84.

[*To be continued*]

Reviews

West Virginia Geological Survey. Marion, Monongalia and Taylor Counties. By RAY V. HENNEN. 1913. Pp. 844, pls. 33, figs. 11, maps 3; Cabell, Wayne and Lincoln Counties. By C. E. KREBS. Pp. 483, pls. 26, figs. 6, maps 9.

The report on the counties of Marion, Monongalia, and Taylor discusses the industrial development, the physiography, the geology, and the mineral resources of this district. These counties are in the northern part of the state in the heart of the Appalachian coal field. They are exceedingly productive of gas and oil and are rich in building stones, glass sands, and clays. To all this must be added great agricultural fertility.

The surface of the region has been eroded to a strikingly mature stage; it is almost entirely a country of steep slopes and very narrow divides. Structurally this district forms part of the eastern flank of the Appalachian Basin. The formations have a general western dip which is modified by local folds. The structure contours on the geological map make it evident that the gas has generally collected in the tops of the anticlines, and that the oil has been found most abundantly upon the flanks of the anticlines. A general section of the strata is as follows:

Upper Carboniferous (Pennsylvanian), 2,300–2,800 feet:
- Dunkard, or Permo-Carboniferous Series, 1,100–1,200 feet
- Monongahela Series, 260–400 feet
- Conemaugh Series, 500–600 feet
- Allegheny Series, 225–350 feet
- Pottsville Series, 250–300 feet

Lower Carboniferous (Mississippian), 450–1,000 feet:
- Mauch Chunk Shales, 40–250 feet
- Greenbrier Limestone, 15–150 feet
- Pocono Sandstones, 400–600 feet

Devonian:
- Catskill Sandstones, 300–500 feet
- Chemung and Hamilton Shales, penetrated in Wheeling deep well nearly 2,000 feet without reaching the Corniferous.

The Dunkard beds are the so-called "Barren Measures," but they carry several seams of coal; these are unimportant commercially but

they are useful for purposes of stratigraphical correlation. The Monongahela series carries the famous "Pittsburgh" coal bed. It is from 8 to 10 feet thick, mined from nearly 60 mines in this one district. The "Sewickley" is generally about 6 feet thick; it is second only to the Pittsburgh coal in economic importance. In the Allegheny series the most valuable coal seams are the Upper and the Lower Kittanning.

The Pocono series consists of alternating beds of sandstones and brown shales. It has been of enormous economic importance, because it contains four oil sands: the Keener, Big Injun, Squaw, and Berea. Oil and gas have been produced from various sands between the Moundsville in middle Conemaugh series down to the Bayard sand near the bottom of the Catskill beds; this is a vertical range of about 1,000 feet.

The report is accompanied by a set of three maps, to make plain the topography, the soils, and the economic geology of the district. The structure contours on the geological map are of especial value and significance in relation to the deposits of coal, gas, and oil.

Cabell, Wayne, and Lincoln counties are in the extreme western part of the state. Their topography is very rough, typically mature. The geological formations dip in a general northerly direction, but they have been somewhat warped by minor folding. The following formations outcrop in these counties, in the order of their appearance, starting from the north: the Dunkard, Monongahela, Conemaugh, Allegheny, and the Kanawha. The removal by erosion of the Monongahela from most of this area, and the irregularity and distribution of the Pittsburgh seam, where the formation still remains, prevent this district from being a great coal-producer. The coal of the Allegheny and Kanawha series is not enough to put these counties very far from the smallest coal-producers in the state. They produce about 100,000 tons per annum as compared with about 4,500,000 tons shipped from the mines of the other three counties reported upon.

The north-central part of these counties is a large oil-producing region; the report suggests that the southern part of the area will be a fruitful field for the oil and gas prospector.

For each of these counties there is a set of three maps which show the agricultural soils, the topography, and the general and economic geology. The maps are made on the scale of one inch to the mile; they have structure contours showing the depth below the surface of the economic deposits and the stratigraphical position of oil and gas accumulations.

T.T.Q.

Water Reptiles of the Past and Present. By S. W. WILLISTON, of the University of Chicago. Chicago: The University of Chicago Press, 1914. Pp. 251, text fig. 131.

It is a deplorable fact, but nevertheless true, that the popular knowledge of present-day reptilian life is very limited. In the opinion of the author of this volume: "In most persons the word reptile incites only feelings of disgust and abhorrence; to many it means a serpent, a cold, gliding, treacherous, and venomous creature shunning sunlight and always ready to poison."

Naturally the public's knowledge of extinct reptiles is much more fragmentary and usually confined to inflated newspaper accounts of "monster" dinosaurs, and a few other forms, possibly. In this volume the author attempts to make the reader familiar with one of the most interesting phases of reptilian life, perhaps, the aquatic forms and the modifications that fitted them for this mode of life. Chaps. i–vi define, in an unpretentious manner, the Reptilia and tell of their occurrence in the rocks and their collection and restoration, their anatomy and classification, their distribution geographically and geologically, and the laws some of them have followed in their adaptation to a life in the water. A classification is given in chap. ii that conforms in the main with the ideas of the more conservative paleontologists. The class is divided into the following orders: the Cotylosauria, Chelonia, Theromorpha, Therapsida, Sauropterygia, Icthyosauria, Squamata, Rhynchocephalia, Parasuchia, Crocodilia, Dinosauria, and Pterosauria. To these the Proganosauria, the Protorosauria, and the Thalattosauria are added and are provisionally given ordinal rank. Chaps. vi–xvi take up in order the Sauropterygia, Anomodontia, Ichthyosauria, Proganosauria, Protorosauria, Squamata, Thalattosauria, Rhynchocephalia, Parasuchia, Crocodilia, and the Chelonia. The various changes these forms have undergone in their adaptation to a life in water, their habits, and, in many cases, interesting bits of history connected with the discovery of the specimens, are told in a fascinating manner.

A few inaccuracies occur in the text and the lettering of some of the figures and some minor additions might be suggested. Among the vertebrate localities given on p. 52 should be included the Pennsylvanian beds from which Case has described vertebrates (*Ann. Carnegie Mus.*, IV [1908]), and on p. 54 might be mentioned the Hallopus beds of Marsh in connection with the Lower Jurassic.

One of the noteworthy features of the book is the large number of excellent illustrations which are, for the most part, the work of the

author. Especially interesting are the life-restorations, some of which are here published for the first time. There is no one better fitted to discuss the subject-matter contained in this book than the author of this volume. In it he gives to the public the benefit of his observations gathered from over forty years of actual experience in the reptilian field. It is rare indeed that a subject is given such an authoritative, scientifically exact treatment combined with a style so thoroughly understandable and interesting to the non-scientific reader. The work is sure to be very popular with the scientist and the general public alike.

M. G. MEHL

The Climatic Factor as Illustrated in Arid America. By ELLSWORTH HUNTINGTON, Assistant Professor of Geography in Yale University, with contributions by CHARLES SCHUCHERT, ANDREW E. DOUGLASS, and CHARLES J. KULLMER. Carnegie Institution of Washington, Publication No. 192, 1914. Pp. vi +341, plates 12, maps 2, text figs. 90.

This volume has bearings which make it important to the geographer, historian, archeologist, meteorologist, and geologist, occupying a field where all these sciences meet, but in this review the volume will be discussed from the geological point of view only.

The purpose of the work is to determine the degree to which climatic changes have taken place in southwestern America during the past 2,000 to 3,000 years. In arid and semi-arid regions the amount of rainfall, as affected by pulsatory changes of climate, becomes most variable and critical.

In addition to the study of the climatic changes shown by the expansion and restriction of ancient peoples in America, as controlled by changes in water-supply or vegetation, the present volume contains two novel lines of attack. The first of these is the use of river terraces as evidences of minor climatic changes occurring within the past few centuries as well as in the more distant past. The second is the measurement of the growth rings of trees. Professor A. E. Douglass gives an introductory chapter on a method of estimating rainfall by the growth of trees. He shows that the rings vary in thickness and correlates the rate of growth with the records of rainfall. Following this, Huntington enters upon a most interesting discussion of the curve of growth of the giant redwoods of California. The data were obtained by careful measurements from stumps and extend back with a large number of trees as much as 2,000 years, with a few trees to 3,000 years. The geological importance of this work is readily seen. As Lyell showed that the present is the key to the past in the crustal history of the earth, similarly the key to the climatic history is to be found in the study of the present climates and

their fluctuations. It is found that the changes, though moderate in amount, are more or less sudden and pulsatory. In the search for causes Huntington consequently assigns crustal movements and changes in atmospheric composition as the broad factors of ultimate control, but the pulsations which appear within the historic record and which extend beyond in the record of moraines left by the oscillations of glaciers, the strands of salt lakes, and terraces made by river action cannot, he argues, be due to these causes. Variations in solar radiation are assigned as the most probable cause. Rhythms measured by tens of miles marked the retreat of the Pleistocene ice sheets; rhythms measured by inches, by feet, and by tens of feet are found in the sediments of many geological formations. In many cases they can hardly be ascribed to crustal causes; they are too many and both too long and too short to fit the precession cycle. Thus the geological record is suggestive that our sun through all of terrestrial history has been a variable star.

The second part of the volume consists of a chapter entitled "The Climates of Geologic Time, and is by Professor Schuchert. There is assembled in thirty pages an account of the various lines of evidence which indicate geologic changes in climate. These are finally correlated in a single chart. The curves of coal-making, limestone-making, aridity, and temperature are given, together with curves showing the movements of the strand line and epochs of diastrophism. While the curves are of course only of qualitative value, they serve to show the variability and the cyclic nature of all these factors through geologic time. This chapter thus gives on a large scale and in distant perspective what the first part of the book gives in minute scale and for the human present.

It would appear that the work of Douglass and Huntington on tree growth opens up a field which deserves further study; a study which should be prosecuted within a few years. In this reconnaissance Huntington has averaged together the measurements of many individual trees. But these have grown under unlike conditions of altitude, slope exposure, and ground water. The averaging of these unlike conditions has tended toward obscuring the amount of short climatic oscillations and the trend of longer changes. An intensive study of stumps selected with respect to these variable controls and an exact dating of special sequences of rings by comparison of stumps would seem to be the next step. But in the meantime decay is blurring year by year this most valuable record.

There are of course in this volume degrees of emphasis and points of view which could be questioned, but within the limits of a short review it would confer a wrong emphasis to single out any point for critical comment, when the book as a whole is a contribution of the first order, in facts, in ideas, and in completeness of presentation. It adds fundamentally to the science which in the future will be named if not now—paleoclimatology.

J. B.